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The new Code of Stratigraphic Nomenclature, prepared by the American Commission on Stratigraphic Nomenclature, appeared in the May 1961 issue of the Bulletin of the American Association of Petroleum Geologists, pages 645–655. Separates are available from the AAPG, Box 979, Tulsa 1, Oklahoma, for 50 cents each, postpaid; 20 per cent discount on lots of 50 or more.

Authors of papers to be submitted to GSA for publication are requested to follow the usage established in the new code.

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Paleosols of Bermuda

Abstract: Sayles recognized five buried soils that stratigraphically separate eolianites and interbedded marine limestones in Bermuda. He considered that the paleosols represent interglacial ages or interstadials. Some soils were considered residual, and some accretionary. Estimates of time of weathering of residual soils were made by computing ratios of insoluble residues in parent eolianite to the overlying whole soil.

Field study shows that island-wide correlation of the paleosols may be tenuous. No available cuts permit direct tracing of soils. Buried soils cannot be dated unless superposition is shown, and only section McGall's Bay shows all five soils superposed; correlation with other sections is difficult.

The morphology, physical and chemical properties, and mineralogy (as determined by X-ray, DTA, and thin sections) of the paleosols show that Sayles' Shore Hills and St. George's soils have considerable soil development and may represent major stratigraphic breaks. They are not residual but have other materials incorporated in their uppermost parts. Thus, Sayles' time estimates are questionable. The Harrington, Signal Hill, and McGall's paleosols have only weak color profiles and slight accumulations of organic matter in their uppermost parts. The latter two soils may represent only local pauses in accretion during accumulation of the later Bermuda dunes.

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INTRODUCTION

Background of Present Study

The Bermuda Islands lie about 753 miles southeast of New York in the Atlantic Ocean. The island chain trends northeast and extends from Long. 64°38′45″ W. to Long. 64°53′45″ W. and Lat. 32°14′30″ N. to Lat. 32°23′50″ N.

The senior author conducted field study during August 1957 under a grant from the Penrose Bequest of The Geological Society of America. A study of the paleosols was conducted at the requests of J Harlen Bretz and

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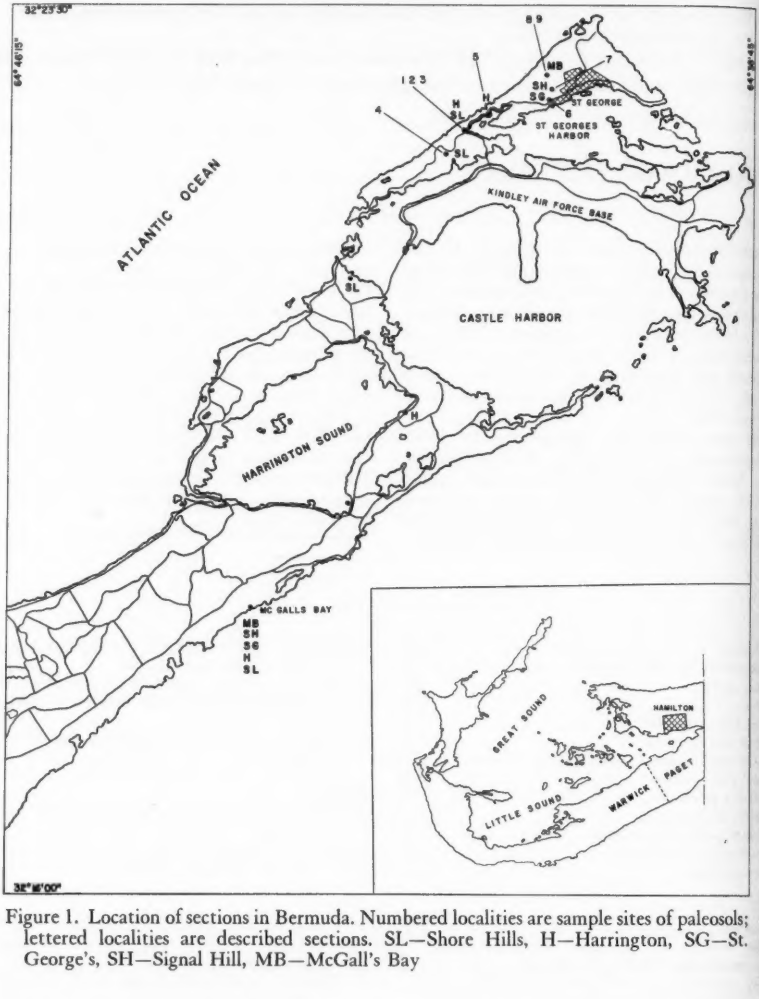

Figure 1. Location of sections in Bermuda. Numbered localities are sample sites of paleosols; lettered localities are described sections. SL—Shore Hills, H—Harrington, SG—St. George's, SH—Signal Hill, MB—McGall's Bay

Heinz A. Lowenstam in conjunction with their geomorphologic and stratigraphic studies.

Because of limited time, the field study was confined to three areas in Bermuda that best display the stratigraphic relations of the sedimentary beds and fossil soils (Sayles, 1931, p. 397). These areas are in St. George's parish, at McGall's Bay, and on the Warwick-Paget line (Fig. 1).

Laboratory studies on the paleosols of Bermuda were conducted during parts of 1958 and 1959 in the Soil Survey Laboratories, Soil Conservation Service, U. S. Department of Agriculture, Beltsville, Maryland, and State College, New Mexico.

Silica was determined by fusion with sodium carbonate and volatilizing with hydrofluoric acid (Kolthoff *et al.*, 1952, p. 384–394). Iron and phosphorus were determined from samples fused with sodium carbonate. Iron was determined from an acid filtrate by titration with potassium dichromate (Hamilton *et al.*, 1947, p. 187–197). Phosphorus was determined from a water filtrate by precipitating it as a pho

ahomolybdate, dissolving in sodium hydroxide, nd titrating the excess base (Horwitz *et al.*, 955, p. 35). Organic carbon was determined y wet digestion with chromic and sulfuric cids and titration with ferrous sulfate (Peech *t al.*, 1947, p. 5–7). Calcium carbonate[1] was etermined by gravimetric loss of carbon ioxide (Richards *et al.*, 1954, p. 105). Alumi- um was determined colorimetrically (Jackson, 958, p. 297–300). The pipette method was sed for mechanical analyses (Kilmer *et al.*, 949, p. 15–24).

The senior author is grateful to The Geo- ogical Society of America for allocation of unds to cover transportation and field ex- penses. Criticism of the manuscript by J Harlen Bretz is appreciated. The writers also appreci- te the assistance of Mrs. Doris P. Gartner in he laboratory studies and V. H. Gledhill's ad- vice on laboratory procedures.

Sayles' Work

STRATIGRAPHY: Sayles (1931, p. 390–397) described the sedimentary rocks that are exposed in the islands as a sequence of eolianites with interbedded marine limestones (Table 1). He defined eolianite as an indurated eolian limestone. The eolianite is described as a white or cream-colored, wind-deposited, more or less consolidated rock composed dominantly of shell fragments, ranging from rounded to angular. Particles range from 0.1 to 2 mm and average 0.5 mm; they are chiefly fragments of pelecypods and some Foraminifera, crabs, and algae. The eolianite is generally cross-bedded.

Sayles (p. 392–394, 430–439) found that several species of extinct land snails, of which the largest and most abundant is *Poecilozonites nelsoni*, occur in the Walsingham eolianite and in the fossil soils, including that in the uppermost part of the Devonshire limestone. No diagnostic faunal assemblages exist above this soil zone.

Sayles (p. 439–455) found no systematic and constant differences in the sedimentary rocks and fossil soils; he therefore concluded that petrology could not be of great service in correlation.

Sayles (p. 391) states: "Due to the eolian nature of the deposits, overlaps and unconformities are abundant. In many sections some of the formations may be completely lacking."

[1] Calcium carbonate as used in this paper is $CaCO_3$ equivalent in per cent and is designated in figures and tables as $CaCO_3$.

A difficulty becomes evident at this point. If there are no diagnostic faunal assemblages, if no systematic differences in petrology exist between beds and soils, if overlaps, unconformities, and absence of formations are abundant, how is a specific formation identified? Presumably if *Poecilozonites nelsoni* is present, the section is somewhere from the Walsingham up through the Harrington. If the snail is absent, the formation could be any one of the four eolianites from the base of the column to the top.

Sayles defined his formations above the Walsingham by relating them to the superposed fossil soils. Without diagnostic faunal assemblages, mineralogic or petrologic characteristics, or other physical features, recognition and classification of specific formations within the stratigraphic column rest upon the proper identification of fossil soils.

FOSSIL SOILS: Sayles (p. 427–428) states:

"Three processes have been active in the production of the soils of Bermuda. Much of the material is obviously residual, the insoluble residue of large amounts of eolianite. Soils of this type are well oxidized and either deep-red or chocolate-brown. A second process is by accretion from above. Fine calcareous sediment blown around by the wind has frequently contributed a large quantity of material to the soils. Such soils are very light-colored, either cream, pink, or gray. Thirdly, the material formed by either of the two processes just outlined may be concentrated in hollows by rain-wash. . . ."

The thickest fossil soil in Bermuda is the red Shore Hills (Sayles, p. 393). It ranges from 4 to 12 feet thick. The Harrington soil is light-colored, as much as 4 feet thick, and composed of wind-blown materials (weathered in some places, fairly fresh in others). Sayles considered it a soil of accretion.

The St. George's soil is deep red or chocolate brown, 2 feet thick; Sayles (p. 395) considered it to be a product of weathering. The younger soils, Signal Hill and McGall's Bay (p. 395–396), are light-colored, locally pink, 2 feet thick, and the products of accretion rather than decay.

Poecilozonites nelsoni serves to distinguish the Harrington and Shore Hills soils (p. 430–439). The Harrington, Signal Hill, and McGall's Bay soils would be difficult to differentiate if they contained no distinctive fossils. The deep-red color, solution pockets, and soil base distinguished the Shore Hills and St. George's soils from the others. Presumably these two soils could be differentiated by the Belmont marine

limestone below the Shore Hills and the Pembroke eolianite beneath the St. George. Stratigraphic formations were identified by the subjacent and superjacent fossil soils, but it is apparent that specific fossil soils cannot be identified readily, and in some cases must be related to subjacent and superjacent formations. Accurate stratigraphic classification and correlation are therefore difficult.

Sayles did not describe the fossil soils in det His analytical data cannot be related to a specific part of a soil, and therefore cannot b compared with data reported here.

TABLE 1. PLEISTOCENE STRATIGRAPHY OF BERMUDA
After Sayles (1931)

Stratigraphic column	Correlation		
Present soil		Postglacial	
Southampton eolianite	Glacial	Wisconsin	Hudsonian
McGall's soil	Interglacial		Recession
Somerset eolianite	Glacial		Quebecan
Signal Hill soil	Interglacial		Recession
Warwick eolianite	Glacial		Manitoban
St. George's soil	Interglacial	Sangamon	
Pembroke eolianite	Glacial	Illinoian	
Harrington soil Devonshire limestone Shore Hills soil Belmont limestone	Interglacial	Yarmouth	
Walsingham eolianite	Glacial	Kansan	

Sayles (p. 442–443) computed the ratios of insoluble residues of the Shore Hills and St. George's soils to the residues of the parent material, marine limestone and eolianite. He concluded that 100 cu ft of limestone was destroyed to yield 1 cu ft of soil. He further estimated (p. 448) that 200,000 years or more were necessary to develop the Shore Hills soil and 150,000 years for the St. George's soil. Comparisons then were made (p. 452–456) with Kay's (1931, p. 425–466) estimates of the duration of the Pleistocene of the North American midcontinent region (*See* Table 1).[2] Flint (1957, p. 357, 363) and Kulp (1953, p. 204–206) have used Sayles' stratigraphy in Bermuda for correlation with the Pleistocene in North America.

It is difficult to compare the results of the present study with the work of Sayles because

[2] The Wisconsin glacial stage as given by Sayles followed Leighton (1931, p. 52). Leighton (1933, p. 168) later changed the names of the substages to Iowan (Manitoban), Tazewell, Cary (both included in Quebecan), and Mankato (Hudsonian).

PALEOSOLS

Shore Hills Paleosol

MORPHOLOGY AND PHYSICAL PROPERTIES

The Shore Hills paleosol varies considerably in its morphology and physical properties (Section 1).

Section 1. Shore Hills (?) paleosol, road cut at intersection .3 mile north of Bermuda Biological Research Station (Fig. 1, No. 1)

Depth (inches)	*Description*
0–62	Very pale-brown (10YR 8/3.5[3] calcareous eolianite; massive; friable to indurated; clastic limestone composed dominantly of rounded shell fragments of fine to coarse sand size; distinct sharp boundary to
62–66 (A_b horizon)	Reddish-yellow (5YR 7/7) loamy sand[4]; single grain structure; friable; calcareous; iron-oxide coatings on grains; gradual boundary to

[3] Munsell soil colors, moist soil

[4] Soil-horizon descriptions follow the U. S. Department of Agriculture's Soil Survey Manual (1951, p. 173–234). The terms "aggregates" or "peds" are used synonymously in this paper to denote individual soil-structure units.

Depth (inches)	*Description*
66–72 (B_{1b} horizon)	Yellowish-red (5YR 4.5/6) sandy loam; generally massive but with weak fine subangular blocky structure; matrix friable and calcareous; with calcareous nodules half an inch to 1 inch in diameter; weak discontinuous iron oxide and possible clay coatings on aggregates; gradual boundary to

grains are very sparse. Although there has been slight structural development in the soil and accumulation of clay in some horizons, the presence of abundant primary weatherable material does not indicate intensive weathering.

This paleosol occurs on a flat to slightly convex summit of the buried surface. The limestone that mantles the paleosol is a bedded

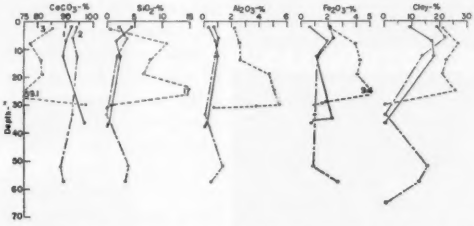

Figure 2. Clay content and chemical composition of moderately developed paleosols. Profile 1—Shore Hills, profile 2—St. George's, profile 3—coalescent Shore Hills and St. George's

72–75 (B_{2b} horizon)	Pink (7.5YR 7/4) loam; moderate fine to medium subangular blocky structure; firm; iron oxide and weak continuous clay coatings on aggregates; interiors of aggregates have single grain structure; calcareous with moderate carbonate cementation; distinct, sharp boundary to
75–76 (D_{rmb} horizon)	Pale-brown (10YR 6/2) dense-banded and laminated capping on bedded and cross-bedded limestone; "soil base" of Sayles; transitional zone half an inch to 1 inch thick to
76–124	White (10YR 8/2) limestone; massive; alternately strongly and weakly indurated; presumably Belmont or Walsingham limestone.

The Shore Hills paleosol in this section is not strongly developed. Clay has accumulated in the B_b horizon as shown by the clay coatings on soil aggregates and the particle-size distribution of the soil (Fig. 2). Total sand content decreases in the A_b, B_{1b}, and B_{2b} horizons from 82.3 to 59.8 to 43.1 per cent. In all horizons the sand is dominantly detrital calcium carbonate grains of rounded fossil-shell fragments. Rounded detrital quartz and black opaque

eolianite, which when traced laterally several hundred feet southwest is surmounted by a moderately developed paleosol. If Sayles' criteria of stratigraphic classification are followed, the upper paleosol must be the St. George's, and the eolianite must be Pembroke. Thus, the Devonshire must be missing from the section, and the Shore Hills paleosol represents weathering from the Belmont to the Pembroke. Bretz (1960, p. 1739–1740) considers the lower limestone to be the Walsingham eolianite, so the weathering interval could be from Walsingham to Pembroke.

A well-developed Shore Hills paleosol, Sayles' type Shore Hills soil (Pl. 1, fig. 1), is detailed in Section 4.

Section 4. Shore Hills paleosol, gravel pit .2 mile northeast of Bermuda Biological Research Station (Fig. 1, No. 4)

Depth (inches)	*Description*
0–14 (IA_b horizon)[5]	Dark-red (2.5YR 3/6) loam; massive; hard; with rounded limestone grains and shell fragments; calcareous; distinct, sharp boundary to

[5] Roman numerals preceding soil horizon designations indicate vertical sequence of different materials (Ruhe and Daniels, 1958, p. 66–69).

Depth (inches)	*Description*
14–20 (IIB_{1b} horizon)	Dark reddish-brown (2.5YR 3/4) clay loam; weak, fine granular; firm; discontinuous clay coatings on aggregates; calcareous; gradual boundary to
20–26 (IIB_{21b} horizon)	Dark reddish-brown (2.5YR 3/4) clay; moderate medium subangular blocky structure; firm; continuous clay coatings on aggregates; upper 3 inches with secondary carbonate, lower 3 inches leached; gradual boundary to
26–46 ($IIIB_{22b}$ horizon)	Dark reddish-brown (2.5YR 3/4) clay; strong medium subangular blocky with macroprismatic structure; firm when dry; plastic when moist; thick continuous clay coatings on aggregates; leached; at 38 inches white secondary carbonate on aggregate faces and in root channels, but matrix leached; gradual boundary to
46–57 ($IIIB_{23b}$ horizon)	Dark-red (2.5YR 3/6) clay; strong medium to fine subangular blocky structure; firm; continuous clay coatings on aggregates; leached; secondary carbonate coatings on aggregate faces and along root tubules; gradual boundary to
57–64 ($IIIB_{3b}$ horizon)	Dark-red (2.5YR 3/6) silty clay; massive; friable when dry; firm when moist; calcareous; distinct, sharp boundary to underlying limestone but clay fillings in solution pipes extend 10–12 feet below surface of limestone
64+ (D_{rb} horizon)	White (10YR 8/2) compact, dense limestone with strong brown (7.5YR 5/4) and reddish-yellow (7.5YR 6/6) iron oxide stain; Belmont limestone.

The Shore Hills paleosol in this section is more strongly developed than in the buried soil in Section 1. Clay has accumulated in the B horizon of this soil (Fig. 3). Undoubtedly, the clay content of the B horizon is greater than reported because complete dispersion was not obtained during laboratory analysis.

Sand contents decrease from 35.6 per cent in the IA_{1b} horizon to 27.5 per cent in the IIB_{1b} horizon and decrease abruptly to 6 per cent in the IIB_{21b} horizon and remain uniform with depth throughout the lower part of the soil profile. Sand in the IA_{1b} horizon is dominantly rounded fragments of limestone and fossil shells. Rounded and subrounded quartz is common. In the IIB_{1b} horizon no limestone and shell fragments are present; rounded and subrounded quartz grains are abundant in this horizon as well as in the subjacent IIB_{21b} horizon. In the $IIIB_{22b}$ horizon and below, quartz grains are sparse. The vertical sequence of differing compositions indicates that three layers of material are involved in this Shore Hills paleosol. The lower material apparently was derived from weathering of the Belmont limestone (Pl. 2, fig. 3). A sediment with abundant quartz was deposited on the lower saprolite, and a second sediment containing carbonate material was deposited above. A deep soil now encompasses all these materials.

Southwest along the face of the quarry, the

PLATE 1. PALEOSOLS, MODERN SOIL, AND McGALL'S BAY SECTION

Scale in feet

Figure 1. Type Shore Hills paleosol; solution pipes extend 10 to 12 feet downward into Belmont limestone; paleosol buried under eolianite to right along quarry face

Figure 2. Harrington paleosol buried under bedded Pembroke eolianite (Mullet Bay section); paleosol is only a weak oxidation profile

Figure 3. Type St. George's paleosol buried under Warwick eolianite; paleosol occupies a depression; paleosolic surface rises up 14 and 21 per cent slopes to the left

Figure 4. Type Signal Hill paleosol; a very weak oxidation profile in Warwick eolianite; buried under Somerset eolianite

Figure 5. Type McGall's paleosol; a very weak oxidation profile in Somerset eolianite; buried under Southampton eolianite

Figure 6. Modern Rendzina with wavy "soil base" of cemented carbonate

Figure 7. McGall's Bay section; from base upward Belmont limestone, Shore Hills paleosol (SL), Devonshire limestone, Harrington paleosol (H), Pembroke eolianite, St. George's paleosol (SG), Warwick eolianite, Signal Hill paleosol (SH), Somerset eolianite, McGall's paleosol (M), Southampton eolianite, modern soil

FIGURE 1

FIGURE 2

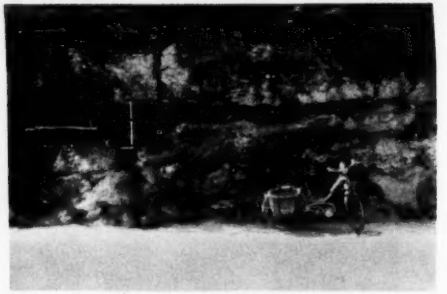

FIGURE 3

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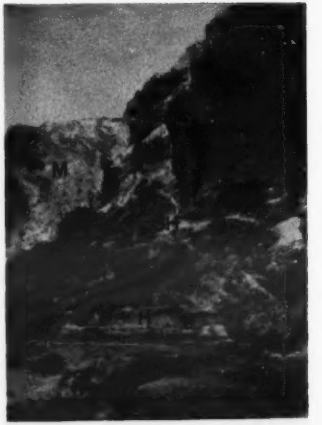

FIGURE 7

PALEOSOLS, MODERN SOIL, AND McGALL'S BAY SECTION
Scale in feet

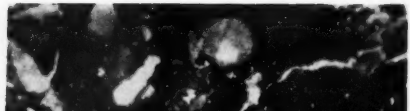

FIGURE 1

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FIGURE 2

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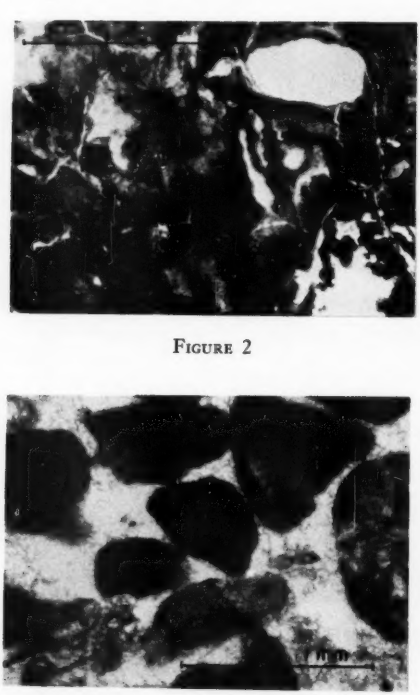

FIGURE 3

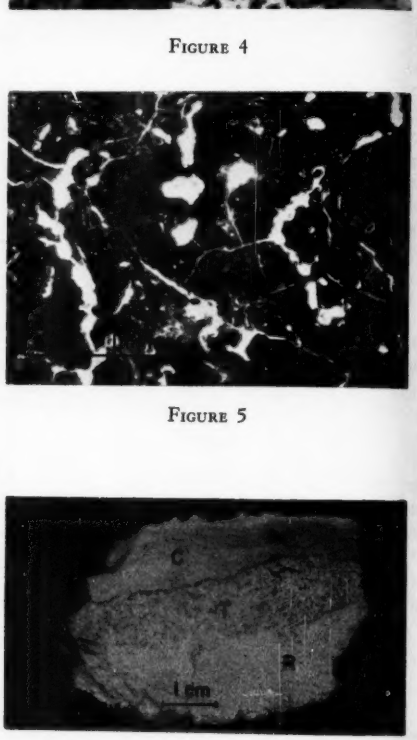

FIGURE 6

MICROMORPHOLOGY OF HORIZONS OF SHORE HILLS AND ST. GEORGE'S PALEOSOLS

Shore Hills paleosol is mantled by a younger limestone. According to Sayles (1931, p. 402), the rock is not the Devonshire marine limestone but is an eolianite of indeterminate age. Bretz (1960, p. 1740), however, believes that the limestone is an early Devonshire dune sand deposited in advance of a rising sea level. The Shore Hills paleosol may thus represent the interval between the Belmont and the Devonshire, or a longer interval between the Belmont and a younger eolianite.

Although this paleosol crops out as a level surface in the quarry face, it probably occupied a depressional position on the surface before burial so that material was washed from adjacent higher slopes into the present upper soil profile. Intensive weathering of soils is not uncommon in depressions with a moist environment (Ruhe, 1956, p. 449). Such soils are generally thicker and more weathered than are those on better aerated higher slopes and summits. The differences in development of the Shore Hills paleosol (Fig. 3, profile 4[6]) in contrast to the Shore Hills paleosol (Fig. 2, profile 1) reflect environmental differences as well as accretion.

MINERALOGY: Sayles (1931, p. 439–444) has described the sand-fraction mineralogy of the fossil soils and eolianites (Table 2). The percentages are averages of the following number of samples of soils and eolianites: Shore Hills (10), Harrington (8), St. George's (11), Signal Hill (7), McGall's (1), and eolianites (9). Sayles observed that perovskite, pyroxene, magnetite, titanite, biotite, and glass occur in the igneous platforms below the sedimentary rocks of Bermuda but thought quartz, orthoclase, garnet, zircon, tourmaline, rutile, muscovite, hornblende, staurolite, and kyanite were erratic because they were not reported in the igneous platform. He (1931, p. 443–444) speculated that these minerals were transported to Bermuda by birds, tropical hurricanes, a jellylike organic substance floating in the ocean, and/or floating sargasso weed.

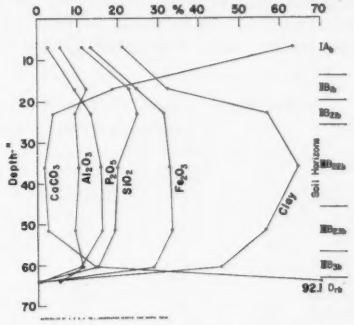

Figure 3. Clay content and chemical composition of well-developed "type" Shore Hills paleosol; profile 4

Clay fractions of profile 4 (Fig. 3) have been analyzed by X-ray and differential thermal analysis. In addition, thin sections of the soil horizons were studied. X-ray patterns consist of very weak reflections which correspond to spacings of 14 and 7.2 A with a band of irregular scatter from about 9 to 4 A. The patterns indicate that the clay contains amorphous material. Small constant amounts of vermiculite and kaolin are present in all horizons. DTA

[6] Soil profile: a vertical section of the soil from the surface through all its horizons into the parent material.

PLATE 2. MICROMORPHOLOGY OF HORIZONS OF SHORE HILLS AND ST. GEORGE'S PALEOSOLS

Figure 1. Photomicrograph of thin section of A horizon of type Shore Hills paleosol; light-gray primary carbonate sand grains in dark clay matrix; an accretionary layer in the soil

Figure 2. Photomicrograph of thin section of B horizon of type Shore Hills paleosol; gelatinous irregularly, finely laminar clay; absence of primary carbonate sand grains

Figure 3. Photomicrograph of thin section of Belmont limestone; parent rock of lower part of type Shore Hills paleosol; rounded sand grains of fossil shells and primary carbonate in calcitic matrix

Figure 4. Photomicrograph of thin section of A horizon of type St. George's paleosol; light-gray sand grains of fossil shells and primary carbonate in dark clay matrix; an accretionary layer in the soil

Figure 5. Photomicrograph of thin section of B horizon of type St. George's paleosol; dark clay matrix with black rounded lateroid aggregates; absence of primary carbonate sand grains

Figure 6. Polished section of "soil base"; C—cap, T—transition, R—rock; occurs at top of rock just below paleo-solum of profile 3; note wavy laminae in cap; result of translocation of iron oxide, alumina, and silica from overlying solum and precipitation in upper part of subjacent rock; *see* Figure 6.

patterns are similar for all horizons and have a broad endotherm at about 330°C which probably is caused by hydrous iron oxides. A fairly broad but definite endotherm occurs at 450–480°C which is caused by a calcium aluminum phosphate, crandallite. Chemical and X-ray-diffraction data support this interpretation.

Thin sections show a mixture of fossil fragments with reddish clayey material and fine secondary calcite in cracks and disseminated throughout the IA_b horizon (Pl. 1, fig. 1). The clayey material is organized in subrounded irregular bodies; some is oriented around fossil fragments. In the IIB_{1b} horizon reddish-brown clayey material contains smaller, more dense aggregates. The clayey material is organized in subrounded irregular aggregates. Some lighter-colored yellowish-brown aggregates contain less granular material and have clay orientation in weakly birefringent films. Secondary calcite fills spaces in the matrix. In addition a colorless mineral, possibly crandallite, with weak birefringence shows extinction characteristics of aggregate crystals. The IIB_{21b} horizon is similar to IIB_{1b}, an unorganized, isotropic clay matrix with a subrounded, blocky aggregate structure. Secondary calcite occurs in cracks and voids. The light-colored aggregates, with a suggestion of oriented clay films, are more abundant.

The $IIIB_{22b}$ horizon (Pl. 2, fig. 2) is distinctly different from the IIB_{21b}. An extremely fine, reddish-brown gelatinous-appearing material has a strong angular crack system indicating colloid movement, shrinkage, and compression. Most of the material shows a faint birefringence with considerable orientation and organization probably due to pressure. A colorless mineral filling cracks and voids has approximately the same birefringence as the material in the matrix but is somewhat better crystallized. The colorless mineral probably is crandallite. The gelatinous-appearing material here looks and reacts like montmorillonite. It swells in water and breaks apart in little rhombohedral slickensided blocks when slightly moist or dry, but X-ray patterns indicate no montmorillonite.

The material in horizon $IIIB_{3b}$ differs structurally from that of $IIIB_{22b}$. It has a fine subrounded irregular aggregate pattern, appears less gelatinous, and has not been affected so much by shrinking and swelling. The matrix is faintly birefringent with weakly birefringent films or layers of clay lining pores and channels.

The thin-section data support the earlier conclusion that accretion of material, including carbonate, fossil shells, and sand grains, has occurred in two upper layers, the IA_b horizon and the IIB_{1b}–IIB_{21b} horizons. These layers overlie a basal layer which includes the $IIIB_{22b}$–$IIIB_{3b}$ horizons.

Table 2. Mineralogy of Sand Fractions of Bermuda Paleosols and Eolianites

After Sayles (1931, p. 421–442); expressed in per cent; T, trace

Mineral	Shore Hills	Harrington	St. George's	Signal Hill	McGall's	Average	Eolianites
Quartz	43	32	25	25	40	32	37
Pyroxene	12	36	25	43	6	27	10
Perovskite	30	16	19	21	6	21	15
Magnetite	6	1	12	9	4	7	5
Orthoclase	T	3	9	1	6	4	18
Glass	T	5	5	0	38	3	6
Titanite	2	5	1	T	..	2	T
Garnet	5	T	T	T	..	1	4
Plagioclase	..	2	2	T	..	1	..
Others	T	T	T	T	..	1	..

CHEMICAL COMPOSITION: Organic carbon (Table 3) has a uniform distribution in the Shore Hills paleosol (profile 1) and may have accumulated in the B_{2b} horizon (0.52 per cent) which occurs just above the soil base.

In the Shore Hills paleosol, profile 4 (Fig. 3), organic carbon has accumulated in the solution pipe fillings (0.96 per cent in $IIIB_{3b}$ horizon) in the Belmont limestone (Table 3) and is uniform in the remaining profile. In both soils maximum organic carbon occurs just above a more impermeable cap on the bedrock, the "soil base," or the bedrock. The more impermeable bedrock surface apparently acted as a catchment for organic material.

Calcium carbonate content (Figs. 2, 3) differentiates the intensity of weathering of the two Shore Hills paleosols. In profile 1 the content is high in the upper part of the solum, somewhat lower in the lower part, and higher in the parent material. Secondary enrichment undoubtedly has occurred in the paleosol by calcium carbonate brought downward in solution from the overlying eolianite. This soil contains 82.3, 59.8, and 43.1 per cent sand composed of fossil-shell fragments in the A_b, B_{1b}, and B_{2b} horizons. Most of the calcium carbonate in the soil probably is primary and indicates that this Shore Hills paleosol is not intensively weathered.

The calcium carbonate content in profile 4 (Fig. 3) is somewhat high in the upper part of the solum, extremely low in the lower part, and high in the underlying rock. Thin sections showed secondary calcite in cracks and voids in all horizons, indicating that secondary enrichment occurred from above. The IA_b horizon

TABLE 3. CHEMICAL COMPOSITION OF MODERATELY AND WELL-DEVELOPED BERMUDA PALEOSOLS

E, eolianite; Ca, cap (soil base); Tr, transition (soil base); R, rock (soil base); A–D, soil horizons

Paleosol and horizon		SiO_2	Fe_2O_3	Al_2O_3	Organic carbon	$CaCO_3$	P_2O_5
Profile 1. *Shore Hills*							
E		.39	.29	.04	.33	96.0	
A_b		4.50	.65	.36	.24	92.1	
B_{1b}		1.71	1.97	.85	.34	90.3	
B_{2b}		2.44	1.12	1.00	.52	89.3	
D_{rmb}	Ca	.30	2.26	.21	..	94.7	
	Tr	.09	.92	.19	..	95.4	
	R	.00	.77	.11	..	97.0	
Profile 2. *St. George's-Shore Hills*							
A_b		2.08	2.12	.74	.33	94.1	
B_b		3.66	2.37	1.06	.35	92.9	
C_b		1.84	1.17	.84	.27	94.5	
D_{rb}		.00	1.00	.15	.31	92.6	
A_{b2}		3.69	.86	1.39	.27	88.5	
B_{b2}		3.34	2.66	.53	.30	89.2	
Profile 3. *St. George's-Shore Hills*							
A_b		.54	2.46	2.22	.40	85.5	
B_{1b}		10.76	3.98	2.60	.48	77.3	
B_{2b}		7.71	4.29	2.65	.34	81.4	
B_{31b}		6.71	4.12	4.73	.34	81.8	
B_{32b}		17.02	9.38	5.09	1.56	59.1	
D_{rmb}	Ca	.77	1.80	5.43		92.7	
	Tr	.09	1.52	3.78		93.6	
	R	.00	1.00	.71		97.5	
Profile 4. *Shore Hills*							
IA_b		11.12	13.16	5.91	.40	63.1	2.7*
IIB_{1b}		22.76	24.54	12.05	.25	18.6	9.1
IIB_{21b}		24.71	31.35	9.21	.44	3.9	13.4
$IIIB_{22b}$		19.95	32.60	10.16	.54	1.9	15.7
$IIIB_{23b}$		19.05	33.09	9.26	.43	2.6	16.0
$IIIB_{3b}$		14.83	28.63	11.06	.96	14.6	10.8
D_{rb}		.36	.60	5.20	.20	92.1	.8
Profile 6. *St. George's*							
E		.56	.83	.10	.60	95.0	
A_b		25.18	9.87	3.54	.78	60.0	
B_{1b}		49.42	12.87	3.43	.62	24.7	
B_{2b}		66.41	14.04	4.25	.60	3.1	
D_{rb}	Ca	4.39	1.52	1.65		92.2	
	Tr	.58	.60	.95		95.3	
	R	.09	.46	.38		96.6	

*P_2O_5 analysis in part by J. H. Caro, U. S. Dept. Agriculture, Agricultural Research Service, Fertilizer Investigations Research Branch

has 35.6 per cent sand composed of fragments of calcareous fossil shells; the source of secondary carbonate is not only the overlying eolianite but also the upper layers of the soil added by accretion.

Iron oxide contents (Figs. 2, 3; Table 3) of the two soils also show differences in degrees of weathering. In profile 1 iron oxide content is low in the upper part of the solum, increases in the lower part, and decreases with depth. An increase occurs in the cap zone of the soil base. Iron oxide has accumulated in the B horizon of the soil and apparently is related to paleosol development. Comparison of the buried soil with the overlying eolianite indicates that enrichment of iron oxide in the paleosol did not come from solutions from the overlying eolianite. Iron oxide content, however, is low (only a few per cent) throughout the profile.

In profile 4 (Fig. 3; Table 3) the iron oxide content is relatively high in the upper part of the solum, increases with depth, and decreases slightly in the lower part of the solum. In the underlying rock the iron oxide content is extremely low. Iron oxide has accumulated in the B horizon of this soil in much greater quantity than profile 1. Profile 4 is much more weathered.

In profile 1 (Fig. 2; Table 3) silica and alumina are low throughout the solum, but increase slightly in the lower part of the solum. In profile 4 (Fig. 3) these compounds are relatively high throughout the solum. Silica is at a maximum in the upper part of the B horizon and reflects the abundance of detrital quartz. With this exception the silica distribution is uniform, but amounts are smaller in the lower part of the solum. Alumina is low in the IA_b horizon but higher and fairly uniform in the rest of the profile.

Profile 4 is unusual with a large content of phosphate (Fig. 3; Table 3) (low in the upper part of the solum, maximum in the B horizon, and decreasing in the lower part of the solum). The underlying rock is extremely low in phosphate. The leaching of guano from the soil surface and precipitation of phosphate as crandallite in the soil has added extraneous phosphate to the profile. Sayles (1931, p. 443–444) noted that Bermuda is on the bird migratory route between the West Indies and Nova Scotia.

The distributions of calcium carbonate, iron oxide, silica, alumina, and phosphate support the interpretations of X-ray-diffraction patterns and differential thermal analysis curves that much of the clay is amorphous and has small amounts of vermiculite and kaolin, and

TABLE 4. MECHANICAL AND CHEMICAL COMPOSITION OF WEAKLY DEVELOPED BERMUDA PALEOSOLS
E, eolianite; A, C, D, soil horizons

Soil and horizon		Sand	Silt and clay	Fe_2O_3	Organic carbon	$CaCO_3$
Profile 5.	*Harrington*					
	E	..	..	.72	.41	94.9
	C_{b1}	96.5	3.5	.74	.38	96.1
	C_{b2}	97.4	2.6	1.66	.36	94.1
	C_{b3}	98.2	1.9	1.29	.30	93.3
	D_{rb}	..	..	.74	.58	92.6
Profile 7.	*Signal Hill*					
	E	..	..	1.29	.44	97.4
	C_{b1}	98.8	1.2	1.12	.37	97.4
	C_{b2}	99.3	.7	.92	.35	97.5
	D_{rb}	..	..	.74	.65	98.1
Profile 8.	*McGall's*					
	E	99.5	.5	.49	.32	99.3
	A_{1b}	96.7	3.3	.60	.49	92.9
	C_b	99.6	.4	.49	.38	94.5
	D_{rb}	99.8	.2	.40	.32	92.9
Profile 9.	*Modern Rendzina*					
	A_1	96.8	3.2	2.23	1.72	92.9
	AC	97.7	2.3	1.52	.91	92.8
	C	99.5	.5	.74	.42	93.9
	D_r	99.7	.3	.60	.42	95.6

that hydrous iron oxides and calcium aluminum phosphate occur in the clay fraction.

Harrington Paleosol

MORPHOLOGY AND PHYSICAL PROPERTIES: The Harrington paleosol (Pl. 1, fig. 3) is quite uniform in its morphology and physical properties. Sayles (1931, p. 394) considered this soil to be one of accretion although "greatly weathered in some places." Section 5 gives the details of this paleosol.

Depth (inches)	*Description*
140–152 (D_{rb} horizon)	Pink (7.5YR 8/4) to white (10YR 8/2) calcareous limestone; horizontal bedding; partially indurated; presumably Devonshire limestone; buried soil over D_{rb} horizon ranges in thickness from 6 to 20 inches.

The limestone in which the Harrington soil is formed appears to be an eolianite rather than a marine limestone. Bretz (1960, p. 1738) states that he and Lowenstam found "roots" of the Shore Hill soil close to tide level at Mullet Bay that are filled with Devonshire marine sand. He believed that the Harrington soil was formed in the upper Devonshire sand and in the first eolian accretions of the overlying Pembroke. He states (1960, p. 1739): "We [Bretz and Lowenstam] found Devonshire marine fossils in its [Harrington soil] lower part and pulmonate snail shells in the upper reddened zone. It is definitely an eolianite deposit throughout."

Bretz with Lowenstam and later Ruhe traced the eolianite overlying the Harrington soil in the cuts along the road toward St. George. A well-developed soil, Sayles' type St. George's (Table 1), is formed in the eolianite. Thus, the eolianite overlying the Harrington soil is accepted as Pembroke (Sayles' definition, 1939, p. 395).

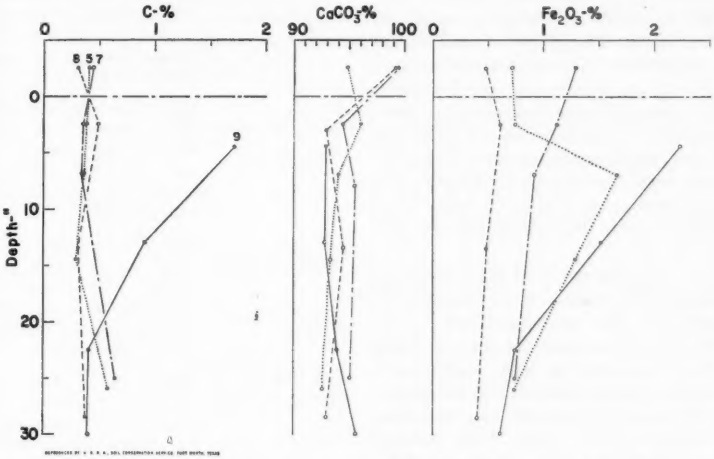

Figure 4. Chemical composition of weakly developed paleosols and modern soil. Profile 5—Harrington, profile 7—Signal Hill, profile 8—McGall's, profile 9—modern Rendzina

Section 5. Harrington paleosol, road cut along Ferry Road at Mullet Bay (Fig. 1, No. 5)

Depth (inches)	*Description*
0–120	White (10YR 8/2) friable calcareous eolianite; horizontal, undulatory bedding; presumably Pembroke eolianite; sharp, distinct boundary to
120–125 (C_b horizon)	Reddish-yellow (7.5YR 8/6) sand; single grain structure; friable; gradual boundary to
125–129 (C_b horizon)	Reddish-yellow (7.5YR 7/6) sand; single grain structure; friable; gradual boundary to
129–140 (C_b horizon)	Reddish-yellow (7.5YR 6/6) sand; single grain structure; friable; sharp, distinct, wavy boundary to

The Harrington paleosol is weakly developed. Sand content (Table 4) is more than 96 per cent throughout the profile. Only minor silt and clay are present. There is no structural aggregation in the soil. Single grain structure, that is, individually preserved carbonate grains, is present in the zone of weathering and in the underlying parent eolianite.

Other Harrington paleosols studied at Harrington Sound and McGall's Bay (Fig. 1, H) have very similar morphologies and physical properties.

MINERALOGY: The mineralogy of the Harrington soil is simple. Individual grains of carbonate rock and shell fragments of sand size are found in all horizons. Very coarse sand is common, but medium and fine sand are dominant. The grains range from subrounded to dominantly rounded. Grain interiors are white, but exteriors are coated reddish yellow. The coatings thicken with depth in the C_b horizon. The coatings are iron oxide (Table 4). A few black opaque grains occur throughout the soil profile.

CHEMICAL COMPOSITION: In comparison to the chemical composition of the moderately developed Shore Hills paleosol (Table 3, profile 1), the Harrington paleosol (Fig. 4; Table 4) has lesser amounts of iron oxide, has a similar organic carbon content, and has greater amounts of calcium carbonate. In comparison to the well-developed Shore Hills paleosol (Table 3, profile 4) chemical constituents differ greatly in the Harrington soil. The latter soil is only a weak color profile in calcium carbonate material. At no place was the soil greatly weathered.

St. George's Paleosol

MORPHOLOGY AND PHYSICAL PROPERTIES: Sayles (1931, p. 429) considered the St. George's soil residual (weathered in place) although he recognized (p. 427) . . . "that as fast as the oxidized material formed [on slopes] it was washed into the hollows by the runoff." The type St. George's soil (Sayles, p. 397) is shown in Section 6. In this cut a reddish clayey paleosol occupies a depression (Pl. 1, fig. 3). The paleosolic surface rises westward in the cut up a 14 per cent slope and more sharply up a 21 per cent slope. The paleosols on these slopes are much lighter colored and lighter textured than the depressional soil.

This paleosol is well developed. Clay has accumulated in the B_b horizon (Fig. 5), and structural aggregates are strongly developed.

Section 6. St. George's paleosol, cut along Ferry Road at west limit of Town of St. George (Fig. 1, No. 6)

Depth (inches)	*Description*
0–96	Very pale-brown (10YR 8/3) calcareous eolianite; bedded; moderately well indurated; presumably Warwick eolianite; modern soil in uppermost part; sharp, distinct boundary to
96–99 (IA_b horizon)	Dark-red (2.5YR 3/6) clay loam; weak fine subangular blocky structure; friable; calcareous with both primary carbonate shell fragments, and secondary carbonate coatings on aggregates; gradual boundary to
99–101 (IIB_{1b} horizon)	Dark reddish-brown (2.5YR 3/4) clay; strong fine subangular blocky; firm; clay coatings on peds; calcareous, with secondary carbonate on ped surfaces; gradual boundary to
101–118 ($IIIB_{2b}$ horizon)	Dark reddish-brown (2.5YR 2/4) clay; strong medium subangular blocky with strong medium prismatic structure; firm; clay coatings on peds; leached; sharp, distinct boundary to
118–169+ ($IIID_{rb}$ horizon)	White (10YR 8/2) calcareous eolianite; horizontal bedding; indurated with pale-brown(10YR 6/3) dense cap one half an inch to 1 inch thick at top of eolianite, soil base, discontinuous in section.

Sand content decreases from 34.2 per cent in the IA_b to 15.3 per cent in the IIB_{1b} to 2.6 per cent in the $IIIB_{2b}$ horizon. Sand in the upper two horizons is dominantly subrounded calcium carbonate fossil-shell fragments. There are sparse rounded detrital quartz and black opaque grains. In the lowest horizon the carbonate fragments are not abundant, but subrounded detrital quartz and black opaque grains are common. The distribution of detrital calcium carbonate sand suggests that at least two accretionary layers are present and have been incorporated in the profile by weathering processes. This soil is not residual; the lower horizon may be weathered from the underlying Pembroke eolianite, but the two upper accretionary layers probably were washed from the adjacent 14 and 21 per cent slopes.

Less weathered and developed St. George's paleosols are described in Section 2. The St. George's soil is formed in Pembroke eolianite

3½ feet thick below the solum. The eolianite in turn overlies the Shore Hills soil. About 50 feet farther southwest in the cut, the St. George's and Shore Hills paleosols merge into one profile. Where the two soils are separated by the thin eolianite the section is as shown in Section 2.

Section 2. St. George's paleosol, road cut .3 mile north of Bermuda Biological Research Station several hundred feet southwest of Section 1 (Fig. 1, No. 2)

Depth (inches)	*Description*
0–66	White (10YR 8/2) fine-grained calcareous eolianite; bedded and cross-bedded; modern soil at top; sharp boundary to
66–70 (A_{b1} horizon)	Reddish-yellow (5YR 6/6) sandy loam; single grain structure; friable; grains rounded limestone fragments and shells; calcareous; gradual boundary to
70–74 (B_{b1} horizon)	Yellowish-red (5YR 5/6) loam; weak fine subangular blocky; peds firm, matrix friable; coatings on aggregates; calcareous; sharp boundary to
74–83 (C_{b1} horizon)	Reddish-yellow (5YR 6/6) sandy loam, calcareous eolianite; mottled with white (10YR 8/2); locally indurated; wavy boundary to
83–116 (D_{b1} horizon)	Very pale-brown (10YR 8/3) sandy loam, calcareous eolianite; generally friable but locally cemented nodules and lenses; sharp boundary to
116–120 (A_{b2} horizon)	Light yellowish-brown (10YR 6/4) loam; mottled with light gray (10YR 7/2); single grain structure; friable; calcareous; gradual boundary to
120–125 (B_{b2} horizon)	Reddish-yellow (7.5YR 6/6) sandy loam; weak fine subangular blocky; firm; coatings on peds; calcareous; sharp boundary to
125–126 (D_{rmb2} horizon)	Pale-brown (10YR 6/3) cap at top of calcareous eolianite; dense; hard; banded; gradual boundary to
126–135+ (D_{rb2} horizon)	Pink (7.5YR 7/4) limestone; bedded; indurated; presumably Walsingham eolianite.

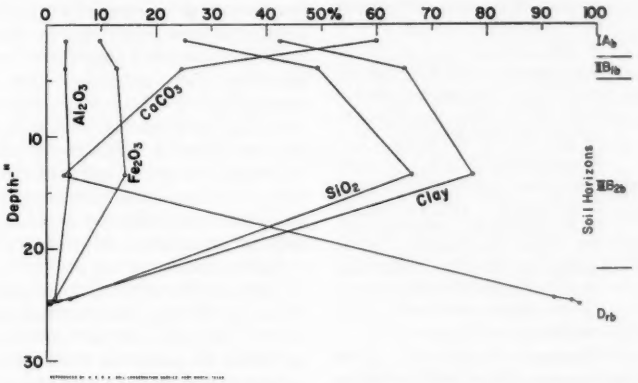

Figure 5. Clay content and chemical composition of well-developed "type" St. George's paleosol; profile 6

The lower buried soil A_{b2}–B_{b2} and underlying limestone D_{rb} can be traced northeastward directly along the face of the cut to Section 1. The lower buried soil is the Shore Hills and is underlain by the Walsingham eolianite of Bretz (1960, p. 1739–40); the upper buried soil A_{b1}–B_{b1} is on a slope of 20 per cent in Pembroke eolianite overlying the Shore Hills paleosol.

Neither buried soil is strongly weathered, although soil development is moderate. In the upper soil total sand contents are 55 and 45 per cent in the solum, and in the lower soil, 47 and 55 per cent. The sand is fresh primary carbonate fossil-shell fragments. Rounded quartz grains are sparse in the upper buried soil but more common in the lower. Clay has accumulated in each of the B_b horizons (Fig. 2); structural aggregation is moderate.

Section 3 gives the detail of the morphology where the two buried soils merge.

Section 3. Shore Hills/St. George's paleosol, road cut .3 mile north of Bermuda Biological Research Station, 50 feet southwest of Section 2 (Fig. 1, No. 3).

Depth (inches)	*Description*
0–109	White (10YR 8/2) calcareous eolianite; bedded and cross-bedded; presumably Warwick eolianite; sharp boundary to
109–114 (IA_b horizon)	Red (2.5YR 4/6) sandy clay loam; single grain structure; friable; calcareous; gradual boundary to
114–120 (IB_{1b} horizon)	Dark-reddish brown (2.5YR 3/4) loam; weak fine subangular blocky; firm; clay coatings on peds; calcareous; gradual boundary to
120–126 (IIB_{2b} horizon)	Reddish-brown (5YR 5/4) loam; moderate fine subangular blocky; firm; clay coatings on peds; calcareous; gradual boundary to
126–130 (IIB_{3b} horizon)	Strong-brown (7.5YR 5/6) loam; moderate fine subangular blocky; firm, compact; clay coatings on peds; calcareous; solution pits 4–6 inches in diameter locally extend 12–18 inches below base of this horizon; filled with strong-brown loam material; sharp boundary to
130–131 (IID_{rmb} horizon)	Pale-brown (10YR 6/3) cap on limestone; dense, banded; half an inch to 1 inch thick; surface wavy; gradual boundary to
131–140+ (IID_{rb} horizon)	White (10YR 8/2) limestone; with strong-brown (7.5YR 5/6) and reddish-yellow (7.5YR 6/6) iron oxide stain; Walsingham eolianite.

This soil is slightly more weathered and somewhat better developed than the two buried soils in Section 2. Textures are heavier; colors have stronger chroma. More clay is present (Fig. 2), but two zones of accumulation are evident. Structural aggregation is moderate.

Fresh primary carbonate fossil-shell fragments are abundant in the upper two horizons which are 55 and 40 per cent sand respectively; the lower horizons have 30–35 per cent sand. The upper two horizons represent an accretionary layer of Pembroke eolianite added to the Shore Hills paleosol in Walsingham eolianite. Two cycles of pedogenesis (Shore Hills and St. George's) are superimposed in one soil profile. This soil is not truly residual soil, nor monogenetic.

A comparison of the three St. George's soils shows that the soil in the depressional area (profile 6) has stronger morphological development than the soils on slopes and summits which are only moderately weathered and developed (profiles 2, 3). Accretionary materials are common in these soils (profiles 3, 6). The superimposition of the second weathering cycle on the soil resulted in more intensive weathering and morphological development than in the soils not subjected to such activity (profiles 2, 3). These complications in pedogenesis make it difficult to estimate age by volume of weathered residue, as Sayles (1931, p. 442–443) proposed.

MINERALOGY: X-ray patterns of the clay fractions of all horizons of the St. George's soil (profile 6) show nonexpanding clays with 14 Å reflections which collapse on heating when saturated with potassium. The mineral is vermiculite. Fairly strong reflections corresponding to a spacing of 7.2 Å indicate the presence of kaolin although the reflections may be in part the second order of the 14 Å reflections. Illite and montmorillonite are not present. Differential thermal analysis patterns are alike for all horizons and indicate the presence of kaolin and about 7–8 per cent gibbsite.

Thin sections of the IA_b horizon of the soil (Pl. 2, fig. 4) show the abundant occurrence of three kinds of carbonate material (primary carbonate fragments of fossil shells, secondary carbonate nodules, and a more coarsely crystalline carbonate that fills the larger pores and channels). The primary carbonate fragments are surrounded by reddish clay that shows aggregate birefringence but only rare thin continuous coatings.

In thin section the IIB_{1b} horizon has a much higher proportion of clay. There are a few fragments of primary carbonate fossil shells. Fine-grained secondary calcite is impregnated throughout the clay matrix; a subangular to subrounded structure is apparent. There are areas of darker and lighter red caused by local variations in amounts of clay and iron oxide and dark-red round dense aggregates that resemble laterite. Some of these aggregates contain gibbsite.

The $IIIB_{2b}$ horizon (Pl. 2, fig. 5) is reddish-brown uniform clay with abundant pores and cracks with considerable organization. Small birefringent scales are common throughout the matrix, and in some places they form continuous patches or networks with the same orienta-

tion. Thin clay coatings are in cracks and oriented around pores. Well-crystallized gibbsite occurs in localized patches between bands of gelatinous-appearing golden-brown very fine clay whose particles have parallel orientation, and in pores and channels in a dark-red unorganized clay matrix.

The X-ray, differential thermal analysis, and thin-section studies show that this St. George's paleosol is strongly weathered and well developed. Clay particles have been oriented within the profile. Increasingly greater contents of detrital calcium carbonate fossil shells in the upper two horizons indicate the addition of two accretionary layers in the soil and give further evidence that this St. George's soil is not truly residual. Analysis of the underlying rock showed no gibbsite, kaolin, or vermiculite.

Little variation appears in the X-ray and DTA patterns in all horizons of the coalescent St. George's-Shore Hills paleosol (profile 3). X-ray patterns have 14 Å and 7.2 Å reflections indicating vermiculite and kaolin. DTA patterns show kaolin and 5–10 per cent gibbsite. Gibbsite increases with depth in the profile.

Thin sections of the IA_b horizon show a loose subrounded ped structure composed of carbonate shell fragments with coatings of reddish-brown clay. The clay shows no oriented birefringence. Clay-particle orientation may be masked optically by the high birefringence of abundant secondary calcite. The IB_{1b} horizon has thicker and more abundant clay coatings surrounding the detrital calcium carbonate fossil-shell fragments. Some of the clay appears to show oriented birefringence, but abundant secondary calcite obscures the pattern. The IIB_{2b} horizon is very similar to the IA_b horizon in thin section, and the IIB_{3b} is very similar to the IB_{1b}. The IIB_{3b} horizon is differentiated by a few dense lateritelike subrounded pellets that contain gibbsite. Secondary calcite obscures possible oriented clay films in the lower two horizons.

The thin sections show not only the added accretionary layer and the two upper horizons of the profile, but also the superimposed weathering cycles. The sequence appears to be an A-B profile overlying an A–B profile. However, in the field the morphology appears as one related sequence of horizons, A over B horizon.

CHEMICAL COMPOSITION: Organic carbon (Table 3) is uniform throughout the St. George's paleosol (profile 6) although there may be a slight accumulation in the IA_b horizon. Distribution of calcium carbonate (Fig. 5) confirms the greater amounts of primary carbonate in the upper two horizons of the profile revealed by mineralogical study. The overlying Warwick eolianite could serve as the source of carbonate for secondary enrichment of the underlying paleosol. Thin-section studies indicate that the upper two horizons of the soil have higher carbonate content because of accretion of primary carbonate and enrichment in secondary carbonate from the overlying eolianite.

Iron oxide increases slightly with depth in the solum (Fig. 5; Table 3). The iron oxide formed during the weathering of the paleosol. There is little if any secondary enrichment from above. Iron oxide content of the overlying eolianite is less than 1 per cent. Alumina increases slightly with depth in the profile, and there has been translocation of this compound. At the soil-base contact with the underlying Pembroke eolianite, the cap zone has 1.65 per cent alumina, the transition 0.95 per cent, and the eolianite 0.38 per cent (Fig. 6). Alumina has concentrated in the soil base.

Silica is extremely high throughout the solum and is at a maximum in the B_{2b} horizon (Fig. 5; Table 3). Silica has also accumulated in the cap of the soil base (Fig. 6) but decreases downward through the transition zone to just a trace in the parent rock.

The presence of abundant amounts of iron oxide, alumina, and silica and their distribution in the profile (Fig. 5) indicate strong development of the soil. The chemical data also support the interpretations of the X-ray-diffraction and DTA patterns.

Organic carbon (Table 3) is uniformly distributed in profile 2, but an accumulation is apparent in the solution pipe filling beneath the solum in profile 3. High carbonate content (Fig. 2) in profiles 2 and 3 is more than 80 per cent with the exception of 59 per cent in the solution pipe filling. Mineralogical data showed this carbonate to be mainly primary carbonate fossil-shell fragments, although some secondary calcite is present. There is no leached horizon with extremely low carbonate content in either profile.

In profile 2 (Fig. 2; Table 3) where two buried soils are separated by thin eolianite, there are two maxima of iron oxide and alumina in each profile. In profile 3, however, where two weathering cycles are superimposed, one maximum of iron oxide and alumina is evident. Apparently the last weathering cycle was

effective in arranging the compounds in the vertical sequence common to one cycle of weathering.

The translocation of iron oxide and alumina in profile 3 is evident when the horizon data of the solum are related to the underlying soil base at the top of the parent rock (Fig. 6; Pl. 2, fig. 6; Table 3). In the soil base cap, iron oxide content is 1.80 per cent, alumina 5.43 per cent, and silica 0.77 per cent. This zone is half an inch thick. A transition zone half an inch thick has an iron oxide content of 1.52 per cent, alumina

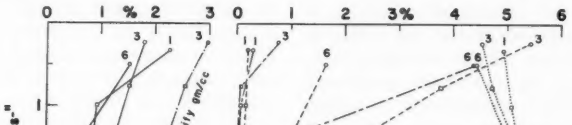

Figure 6. Physical and chemical characteristics of soil base. Profile 1—Shore Hills, profile 3—coalescent Shore Hills and St. George's, profile 6—St. George's

3.78 per cent, and silica 0.09 per cent. The underlying parent limestock has 1.00 per cent iron oxide, 0.71 per cent alumina, and no measurable silica. The densities of these zones show that materials have been added from above. The parent rock has a density of 1.92 gm/cc, the overlying transition 2.54 gm/cc, and the overlying cap 2.96 gm/cc. The cap is just beneath the base of the solum. The soil base is a thin horizon of iron oxide, alumina, and silica accumulations that have been transported from the overlying soil and concentrated in the upper part of the subjacent parent rock.

A comparison of the chemical data (Table 3) of the St. George's paleosols of the poorly drained depression (profile 6) with the polygenetic (profile 3) and the monogenetic (profile 2) soil reaffirms the conclusion reached through morphologic and mineralogic studies that the intensity of weathering and soil development decreases in the order given. Sayles (1931, p. 442–443) did not observe these complexities in the soil in correlating the St. George's paleosol with the continent and in estimating its age by degree of soil development. Thus, his conclusions are questionable.

Signal Hill Paleosol

MORPHOLOGY AND PHYSICAL PROPERTIES: Sayles (1931, p. 397–398) considered St. George's parish the type locality of the fossil soils of Bermuda. He traced the stratigraphic relationships of the eolianites and buried soils directly from the type St. George's soil (profile 6 of this paper) to the top of Khyber Pass where the type Signal Hill soil is present. Sayles stated: "Four of the five soils are exposed in such relations as to preclude any possible doubt that each soil represents a distinct and separate horizon." Bretz and Lowenstam (Bretz, 1960, p. 1738) and later Ruhe attempted to trace the eolianites and soils that Sayles described but could not confirm Sayles' correlation.

Sayles' type Signal Hill soil is present in the pass, however (Pl. 1D), and is exposed in a vertical road cut (Section 7).

Section 7. Signal Hill paleosol, Khyber Pass, St. George's parish (Fig. 1, No. 7)

Depth (inches)	*Description*
0–51	White (10YR 8/2) calcareous eolianite; friable to moderately indurated; upper beds horizontal, lower, cross-bedded, dipping northwest; presumably Somerset eolianite; sharp boundary to
51–56 (C_b horizon)	Very pale-brown (10YR 8/3.5) sand; single grain structure; friable; calcareous; gradual boundary to

Depth (inches)	*Description*
56–66 (C_b horizon)	Very pale-brown (10YR 8/3) sand; single grain structure; friable; calcareous; gradual boundary to
66–138+ (D_{rb} horizon)	White (10YR 8/2) calcareous eolianite; friable to moderately indurated; cross-bedded, dipping southeast; presumably Warwick eolianite.

Morphologically there is little evidence of weathering in this buried soil. There is very slight coloration (only 1 to 1½ units of chroma). There is no clay formation detectable. The C_b horizons have 98.8 and 99.3 per cent sand (Table 4). There is no structural aggregation, and the only horizonation is based upon very slight color changes.

MINERALOGY: All horizons in the section contain individual grains of primary calcium carbonate shell fragments. All grains are subrounded to dominantly rounded. The grains range from very coarse to very fine sand with medium sand dominant. There are a few black opaque grains. In the C_b horizons the grains have very weak, very pale-brown coatings on the surface. Interiors of the grains are white. The grains in all horizons appear fresh and unweathered.

CHEMICAL COMPOSITION: All horizons have more than 94 per cent calcium carbonate (Fig. 4; Table 4). Organic carbon is less than 1 per cent and is uniformly distributed throughout the section. Iron oxide decreases uniformly from slightly more than 1 per cent in the overlying Somerset eolianite to slightly less than 1 per cent in the underlying Warwick eolianite. The very pale-brown coatings on the grains in the Signal Hill soil probably are iron oxide.

The chemistry, mineralogy, and morphology of this soil show that there has been little weathering except for some release of iron oxide.

Sayles (1931, p. 396) considered the Signal Hill soil to be one of accretion. The soil is essentially raw eolianite.

McGall's Paleosol

MORPHOLOGY AND PHYSICAL PROPERTIES: Sayles (1931, p. 398) stated that the McGall's soil is exposed on the St. George's Hotel golf course about 200 yards northeast of the Signal Hill type locality (Section 8). A shallow valley separates this exposure from Signal Hill. Bretz with Lowenstam (Bretz, 1960, p. 1739), and later Ruhe, agreed that no stratigraphic relationships can be determined between these localities. Bretz stated: "Since no order of superposition exists, this McGall's soil could be judged as earlier or later than, or contemporaneous with, the Signal Hill type exposure, at the whim of the observer."

Section 8. McGall's paleosol, sand trap of St. George's Hotel golf course (Fig. 1, No. 8)

Depth (inches)	*Description*
0–180	Very pale-brown (10YR 8/3) calcareous eolianite; presumably Southampton; friable; cross-bedded, dipping south; lower 9 inches massive and friable; modern soil at top; sharp boundary to
180–186 (A_{1b} horizon)	Light-gray (10YR 7/2) sand; single grain structure; friable; calcareous; gradual boundary to
186–201 (AC_b horizon)	Very pale-brown (10YR 8/3) sand; single grain structure; friable; calcareous; gradual boundary to
201+ (C_b horizon)	White (10YR 8/2) sand; single grain structure; presumably Somerset eolianite.

Morphologically there is little evidence of weathering in this buried soil. There appears to be a slight accumulation of organic matter in the A_{1b} horizon and a slight coloration due to slight oxidation in the AC horizon. The magnitude of color change is only one unit of chroma. There is no formation of clay. Sand contents (Table 4) down through the section are 97 or more per cent. There is no structural aggregation, and the only horizonation is based upon slight changes in color.

MINERALOGY: All horizons in the section are composed of grains of calcium carbonate shell fragments that are subrounded to dominantly rounded. The grains range from very coarse to very fine sand with medium sand dominant. All grains appear fresh and unweathered. In the A_{1b} horizon there are black, thin, sparse carbonaceous coatings on the grains. In the AC_b horizon there are thin, very pale-grown iron oxide coatings on the grains. Little weathering is evident.

CHEMICAL COMPOSITION: The slight accumulation of organic carbon (Fig. 4; Table 4) in the A_b horizon of the soil decreases with depth in the profile. Calcium carbonate content is high throughout the profile. There is slight accumulation of iron oxide in the A_b horizon which decreases with depth. The conclusion that little weathering has been effective is supported by

chemical data deduced from the morphology and mineralogy of the soil. There has been only slight accumulation of organic matter in the A_b and AC_b horizons. Sayles considered McGall's soil to be accretionary.

Modern Rendzina

MORPHOLOGY AND PHYSICAL PROPERTIES: For comparison a modern soil was studied, and samples were collected at the same locality as the McGall's soil (Section 9). The soil occurs on a 9 per cent slope.

Section 9. Modern Rendzina sand trap on St. George's Hotel golf course (Fig. 1, No. 9)

Depth (inches)	*Description*
0–9 (A_1 horizon)	Very dark-gray (10YR 3/1) sand; weak fine granular structure; friable; calcareous; heavy root mat; gradual boundary to
9–17 (AC horizon)	Light brownish-gray (10YR 6/2) and grayish-brown (10YR 5/2) sand; single grain structure; friable; calcareous; roots abundant; gradual boundary to
17–28 (C horizon)	Very pale-brown (10YR 8/4) sand; single grain structure; friable; calcareous; depth of root penetration at base of this horizon; gradual boundary to
28–65+	Very pale-brown (10YR 8/3) sand; single grain; friable; calcareous eolianite presumably Southampton.

Morphologically this soil shows stronger development than the Harrington, Signal Hill, or McGall's paleosols. There are greater accumulations of organic matter in the A horizon, stronger chromas in the profile, and evidence of carbonate leaching. A soil base of indurated secondary carbonate a quarter to half an inch thick commonly occurs in the lower part of the soil (Pl. 1, fig. 6). Although sand content (Table 4) is high (more than 95 per cent throughout the profile) it is lowest in the A horizon and progressively increases with depth. Such distribution may reflect the leaching in the upper part of the soil.

MINERALOGY: Individual grains of primary carbonate fragments of fossil shells occur throughout the profile. The grains range from very coarse sand to very fine sand, with medium sand dominant. All grains are subrounded to dominantly rounded. In the A_1 horizon the grains have very dark-gray coatings of organic matter. Interiors of grains are white. In the AC horizon the grains have very dark gray coatings of organic matter, and some have pale-brown coatings of iron oxide. In the C horizon the coatings are pale-brown iron oxide. Carbonate-cemented aggregates are common in the C horizon, which indicates some leaching and reprecipitation of carbonate in the soil.

CHEMICAL COMPOSITION: Organic carbon and iron oxide (Fig. 4; Table 4) are at a maximum in the A_1 horizon of the soil and progressively decrease with depth. The organic carbon and iron oxide contents are much greater in comparison to the Harrington, Signal Hill, and McGall's paleosols. There is no orderly arrangement of these compounds in the paleosols as in the modern Rendzina. Carbonate is at a minimum in the A_1 horizon and progressively increases with depth. The orderly arrangement of carbonate may be due not only to the addition of organic carbon and iron oxide in the upper horizons of the soil, but also to the leaching of carbonate in the upper horizons and reprecipitation at depth. The contents and distribution of the compounds in the modern Rendzina show that it is a more developed and more weathered soil than the paleosols.

Characteristics of Paleosols

The Harrington, Signal Hill, and McGall's paleosols are Regosols, consisting of unconsolidated rock in which few or no clearly expressed soil characteristics have developed (Thorp and Smith, 1949, p. 120). The type Shore Hills and St. George's paleosols are terra rossas. (Terra rossa is a poorly defined term for red soils overlying limestone, but not necessarily derived from it, Robinson, 1949, p. 423–427; Kubiena, 1953, p. 214–219.)

The relative status of the Bermuda soils can be shown by a listing from most developed and weathered to least: depressional Shore Hills (4), St. George's (6); polygenetic St. George's/Shore Hills (3); Shore Hills (1), St. George's (2), Shore Hills (2); modern soil (9); Harrington (5), Signal Hill (7), McGall's (8).

Of all the soils studied and analyzed, the Shore Hills (profile 4) and St. George's (profile 6) paleosols that formed in depressions are the most developed and weathered. They have the greatest clay content and leached subsoil horizons and probably had low base status (Figs. 3, 5). They have high silica and sesquioxide contents (Table 3). Both soils are not residual because they have accretion of extraneous material in their sola. Both the Shore Hills and the St. George's have had two layers added; of

the two soils, the Shore Hills is more weathered.

The polygenetic St. George's-Shore Hills paleosol (profile 3) is next in sequence. This soil formed in the first cycle (Shore Hills) on a level summit, which became, after accumulation of a Pembroke dune, a depression during the St. George's cycle of weathering. The soil has smaller amounts of clay (Fig. 2), silica, and sesquioxides (Table 3) than do profiles 4 and 6. In addition to two cycles of weather, this soil has an accretionary layer in its upper part.

The Shore Hills paleosol (profile 1) and the St. George's-Shore Hills (profile 2) are next in the order and have the least amount of clay (Fig. 2), silica, and sesquioxides.

The modern soil is next in order of weathering. Very little clay has formed in the profile (Table 4). Organic matter and iron oxide have accumulated in the upper part of the soil (Fig. 4). A slight amount of carbonate leaching has taken place. This soil has weathered and developed more than the Regosols (the Harrington, Signal Hill, and McGall's paleosols).

These latter soils have had little clay formed (Table 4) and have accumulated little organic matter and iron oxide. They are no more than slightly oxidized raw eolianite.

Significance of Paleosols

Sayles (1931, p. 442–444) computed the ratios of insoluble residues in the terra rossas and their parent rock and concluded that 100 cubic feet of limestone was destroyed to give 1 cubic foot of soil; he assumed that the insoluble materials in the soil are derived from the limestone.

Sayles (1931, p. 443) recognized that soils on hill slopes lost much material due to rain wash, and in depressions large amounts of material had been washed in from surrounding slopes. He concluded that a red soil developed on a flat surface is a strictly residual soil and based his calculations on such a soil. This last assumption is not necessarily true. For example, the type Shore Hills paleosol (profile 4) occurs on an essentially flat surface, yet there are two accretionary layers in the upper part of this soil. The polygenetic St. George's-Shore Hills paleosol (profile 3) occurs on a flat Shore Hills paleosolic surface but has two cycles of pedogenesis in addition to an accretionary layer in its upper part.

A serious objection to the residue assumption is that additions made to a soil during pedogenesis (organic carbon from vegetation growth, sesquioxides, clay minerals deposited by wind and water, and phosphate leached from guano, profile 4) normally in weak acid digestion will remain as insoluble residue. Thus, the ratio of soil residue to rock will be distorted. Such calculations imply that only subtractions of material occur during pedogenesis, that is, carbonate is removed and insolubles accumulate. The additions are ignored.

Because of the complications in the paleosols and Sayles' erroneous assumptions, his weathering estimates cannot be accepted. (Fairbridge and Teichert, 1953, p. 79, previously challenged Sayles' estimates.) The inherent error in Sayles' computation was compounded when he (1931, p. 428–429) resorted to time estimates of weathering. He estimated that 50,000–60,000 years were required to reduce 100 feet of rock to 1 foot of soil. Average-thickness ratios of Shore Hills and St. George's soils then became factors for computation of time. Average for Shore Hills time was 200,000 ± 50,000 years, and for St. George's 150,000 ± 50,000 years.

The Regosols were estimated at "a few thousand or tens of thousands of years each, and post-eolianite time 25,000 years."

From these values Sayles (1931, p. 452–456) correlated his paleosol times with those of Kay (1931) in the midcontinent region of North America. Sayles accepted Kay's values although the paleosols of Bermuda were formed in rock of almost 100 per cent carbonate whereas the glacial drifts of the mid-continent have carbonate contents of 10–20 per cent. (See Flint, 1949, p. 297–303, for review of leaching of carbonate as basis for age correlation.)

Sayles (1931, p. 397–404) selected St. George's parish as the type area of the fossil soils of Bermuda. Other areas would have been preferable. Despite his statement that "Four of the five soils are exposed in such relations as to preclude any possible doubt that each soil represents a distinct and separate horizon," the stratigraphic relationships of the limestones and buried soils are not easily determined or proved. Bretz (1960, p. 1738–39) with Lowenstam and later Ruhe could not trace the horizons with any degree of certainty.

The best section in Bermuda is at McGall's Bay, but Sayles (1931, p. 406–408) did not discuss it in any detail. In one sea cliff at McGall's Bay all of the paleosols are exposed (Pl. 1, fig. 7). The best part of the section is at the southwest end of the cliff. Here the Shore Hills paleosol occurs on the top of the marine bench about 5 feet above tide. Most of the clay of the soil has been removed by wave activity,

but many solution pipes with heavy red clay fillings remain. The soil is on hard crystalline limestone.

Several feet of indurated bedded sand overlies the Shore Hills paleosol. A zone 19 inches thick in the uppermost part is of very pale-brown (10YR 8/3) sand that is massive but with single grain structure and friable. This zone is the Harrington paleosol and is no more than a weakly oxidized profile. It is, however, a regosolic soil and must be recognized as a stratigraphic break.

Ten feet of cross-bedded eolianite overlie the Harrington paleosol. The St. George's soil is well-developed in the uppermost part of the eolianite. The upper 4 inches are yellowish-red (5YR 5/6) sand that is massive but with single grain structure and friable. This is the A_b horizon of the paleosol. A subjacent horizon 6 inches thick is reddish-brown (5YR 4/4) loam with weak, fine subangular blocky structure and with weak clay coatings on the soil peds. The horizon is firm and is the B_{1b}. The next subjacent 10 inches is reddish-brown (10YR 4/4) clay loam with moderate fine subangular blocky structure and strong, abundant clay coatings on the soil peds. This horizon is firm and is the B_{2b}. A soil base occurs beneath the B_{2b} horizon in the top of the subjacent eolianite.

Overlying the St. George's paleosol is an eolianite 5–5½ feet thick. A zone 15 inches thick in the uppermost part of the eolianite is reddish-yellow (5YR 6/6) sand, massive but with single grain structure and friable. This zone is the Signal Hill paleosol. The soil has redder hue and stronger chroma than any of the so-called accretionary soils of Bermuda, which suggests more intensive oxidation. The reddish-yellow Signal Hill paleosol certainly is oxidized more than the lower lying very pale-brown Harrington. There is no morphological development other than coloration in the Signal Hill soil but this regosolic soil represents a stratigraphic break in the section.

The Signal Hill paleosol is overlain by 10–15 feet of white, cross-bedded eolianite. In the uppermost part of the eolianite is a zone 29 inches thick of very pale-brown (10YR 8/3) sand that is massive but with single grain structure and friable. This is the McGall's paleosol (type of Sayles) but is no more than a weakly oxidized profile in raw eolianite sand (Pl. 1, fig. 5). It is, however, a regosolic soil and a recognizable stratigraphic break. The color of the McGall's soil is the same as that of the Harrington. The zone of coloration is thicker than in the Harrington, suggesting deeper oxidation. The soil is less oxidized than the Signal Hill soil lower in the section. The McGall's soil is overlain by 10 feet of white, cross-bedded eolianite that dips west.

The limestones and paleosols of the McGall's Bay section apparently substantiate Sayles' stratigraphic column of Bermuda. From the base upward are Belmont limestone, Shore Hills soil, Devonshire sand, Harrington soil, Pembroke eolianite, St. George's soil, Warwick eolianite, Signal Hill soil, Somerset eolianite, McGall's soil, Southampton eolianite, and modern soil.

Island-wise correlation of the fossil soils, however, is another question. Internal characteristics of the soils cannot be used for identification and correlation. In the moderately to well-developed paleosols (Shore Hills and St. George's), variability of characteristics prevent identification of a soil as one or the other. The same conclusion can be drawn for the weakly developed paleosols (Harrington, Signal Hill, and McGall's). For example, one exposed paleosol with either a moderately or well-developed soil with textural B horizon, or a weakly developed soil lacking horizonation could not be identified as Shore Hills or St. George's in the first case, or Harrington, Signal Hill, or McGall's in the second. Only where stratigraphic superposition is evident or where direct tracing is possible can correlations be made with any degree of certainty.

However, the Shore Hills and St. George's paleosols represent major stratigraphic breaks in the Bermuda section. The Harrington, Signal Hill, and McGall's paleosols represent more minor breaks, or the latter two possibly only local pauses in accretion during accumulation of the later Bermuda dunes.

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Structural Geology of the Beartooth Mountains, Montana and Wyoming

Abstract: The Beartooth Mountains—an 80 by 40 mile elevated crustal block of Precambrian crystalline rocks—are in the Middle Rocky Mountains east of the great north-south belt of overthrusts. The Beartooth Mountains are bounded on the east by the Bighorn Basin (average structural relief, 15,000 feet); on the northeast and north by the Nye-Bowler lineament (along which movement has been dominantly left-lateral); and on the north by the Crazy Mountain syncline (average structural relief of at least 10,000 feet). The southwest side (Yellowstone-Absaroka Mountains) is mostly covered by thick Tertiary volcanic rocks.

Detailed mapping, supplemented by results of earlier workers, provides an integrated geologic picture of the tectonic development of the block. Much of the tectonic history is recorded in the structures developed in the Paleozoic and Mesozoic sedimentary rocks exposed around the edge of the mountains. The sedimentary rocks played a relatively passive role in the dynamic rise of the block during Laramide time, as they draped, folded, broke, and slid along the edge of the block.

A thrust plane(s) dipping west-southwest is the major structural feature between the Clarks Fork of the Yellowstone and Nye, Montana. Near Red Lodge numerous tear faults displace the upper plate of the thrust as much as 10,000 feet. From Nye to Livingston, Montana, the major bounding structure is a north-northeast-dipping thrust plane(s) terminating in several left-lateral tear faults. Near Gardiner a high-angle thrust plane dipping northeast bounds part of the southwest Beartooth.

Several structural mechanisms apparently operated during deformation: (1) Uplifting, tilting, and depressing of large crustal blocks. These commonly form ramps from the Beartooth massif to adjacent basins; during uplift smaller blocks were rotated in the opposite sense. (2) Vertical raising of the northwest and southeast corners. Here, the general lack of horizontal displacement relative to the adjacent basins limits the total horizontal displacement of the block. (3) Thrust faulting. Most of the frontal thrusts show moderate dips. Imbrication is developed particularly at the unconfined northeast corner. The major thrust steepens with structural depth to the south. The horizontal thrust displacement is not directly measurable at most places but is limited by the lack of horizontal movement at the two fixed corners of the block. (4) Tear faulting. Several major tear faults (displacements up to 10,000 feet) bound large keystone-shaped blocks near Red Lodge. (5) Lateral shearing. During the late stage of uplift, horizontal movement toward the unconfined Red Lodge corner created major lateral shears, associated with imbrication, along the north and east sides, with folding along vertical axes and consequent shortening of the sedimentary section.

Tectonic development is attributed to (1) horizontal compressive forces, responsible for early crustal block movements and subcrustal plastic shifting, and (2) vertical forces of fluid pressures from the continued transfer of subcrustal material which probably helped raise the block. Uplift ceased when one or both of these forces terminated.

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INTRODUCTION

General Setting

The Beartooth Mountains are in south-central Montana and northwestern Wyoming; the center of the range is northeast of Yellowstone National Park (Fig. 1). The Beartooth Mountains constitute a topographic and structural block[1] of the earth's crust about 80 miles long and 35–40 miles wide, trending northwest. Livingston, Gardiner, and Red Lodge, Montana, are, respectively, at the northwest, southwest, and northeast corners of the roughly rectangular uplift (Fig. 2). A sharp corner at the mouth of the Clarks Fork Canyon of the Yellowstone River in Wyoming marks a partial southeastern terminus, although possibly the true southeastern corner is near Cody, Wyoming (Pl. 1).

The name Absaroka Mountains has been used to denote a large part of the Beartooth block as defined above. The authors think this unsuitable in terms of the geology of the region. In this paper the name Absaroka refers only to the widespread eroded igneous, volcanic, and sedimentary rocks, most of them horizontal, that cover a part of the southern part of the Beartooth uplift, and that extend southward to cover the Washakie Range and the western part of the Owl Creek uplift. The authors consider the name Snowy Range, which some authors have applied to the northwestern part of the Beartooth Mountains near Livingston, a confusing and unnecessary complication and ignore its connotation as a separate mountain mass.

Topographically, the Beartooth Mountains rise abruptly 4000–5000 feet from the Great Plains along the east and northeast. The 12,000-foot crest of the range lies close to this frontal edge. Two surfaces, the lower one extensive and well developed, slope gently southwestward from the crest to a place where they are buried by the water-laid volcanic rocks of the rugged Absaroka Mountains and Yellowstone Plateau. Thus the true southwestern boundary of the Beartooth massif is covered.

The broad alluvium-filled valley of the Yellowstone River along the western border of the Beartooth Mountains from Gardiner to Livingston is about 5000 feet below the Beartooth Mountains. The Yellowstone is the master stream of the region, and into it flow a series of youthful streams that drain the Beartooth Mountains along the northeast, including from west to east Mission, West Boulder, Main Boulder, East Boulder, Stillwater, Fishtail, West and East Rosebud, and Rock creeks (Pl. 1). The Clarks Fork of the Yellowstone River flows southeast from near Cooke City, Montana, to its canyon mouth about 25 miles north-

[1] In this paper "mountains" has a geographic or topographic meaning; "block" connotes a well-defined structural unit of the crust; and "massif" connotes an uplifted mass of the crust including basement rocks.

west of Cody, Wyoming (Pl. 2, fig. 1), generally dividing the volcanic rocks on the south from the Precambrian crystalline rocks of the Beartooth Mountains on the north. After its exit from the mountains, the Clarks Fork flows northeastward in the Bighorn Basin toward its junction with the Yellowstone River.

Investigations in the Beartooth Region

During the summer of 1950 the senior author investigated some of the broad structural relationships of the Beartooth Mountains with surrounding areas. Stimulated by Prof. W. T. Thom, Jr., of Princeton University, he became interested in studying the structural geology around the perimeter of the Beartooth massif in an attempt to determine the time and nature of deformation of the entire block with relation to adjacent areas.

During 1954 the senior author designated, from reconnaissance studies, three critical areas for detailed study, and initial mapping was begun. These areas encompassed all the mountain front between the Clarks Fork Canyon, Wyoming, to West Fishtail Creek near Nye, Montana, and between the East Boulder and Boulder rivers southeast of Livingston, Montana. During 1955 and 1956 the three authors completed detailed mapping in the chosen areas as well as careful field checking in areas that John S. Vhay (1934, Ph.D. thesis, Princeton Univ.) and Richards (1957) had studied. Further studies around the perimeter of the massif by Foose and Wise have taken short periods of the 1957 and 1958 field seasons.

Objectives and Scope

The primary objectives of this investigation have been to:

(1) Map the structural relationships along a major part of the east and northeast sides of the Beartooth Mountains.

(2) Incorporate with the present studies the best and most recent mapping of others in adjacent areas and to compile a geologic map of the structural relationships around the perimeter of the Beartooth block, as well as a portion of the Precambrian core.

(3) Relate the structural patterns of the Beartooth mountain front to adjacent structural units.

(4) Determine the chronology of geologic events particularly as they pertain to the Beartooth Mountains.

(5) Interpret the nature of deformation of the Beartooth block by relating the facts to each other and separating these insofar as possible from speculations.

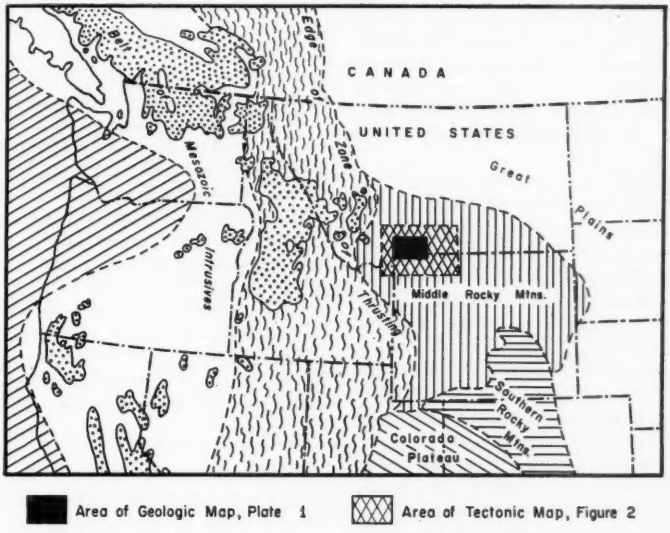

Figure 1. Location of the Beartooth block

Prior Work

Since the organization of the Yellowstone-Bighorn Research Association in 1930, many individuals and teams have attacked the problems of Laramide orogeny in the Middle Rocky Mountains (Pl. 1, index).

Among those who have contributed geologic details along the perimeter of the Beartooth block and of adjacent areas are Hughes (1933; 1932, Ph.D. thesis, Johns Hopkins Univ.), Cloos and Cloos (1934), Wilson (1934), J. S. Vhay (1934, Ph.D. thesis, Princeton Univ.), Dorf (1934), Perry (1935), Lammers (1937), Rouse *et al.* (1937), D. C. Skeels (1939, Ph.D. thesis, Princeton Univ.), Peoples and Howland (1940), Howland, Garrels, and Jones (1949), Howland (1955), Richards (1957), C. W. Brown (1957, Ph.D. thesis, Princeton Univ.), Pierce (1957), and Hess (1960).

Details within the crystalline core of the Beartooth massif have been provided by the works of Iddings and Weed (1894), Hague (1899), Lovering (1929), J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.), Parsons (1939; 1958), Eckelmann and Poldervaart (1957), Poldervaart and Bentley (1958), Spencer (1959), and Harris (1959). Robert Butler, Clarence Casella, Martin Prinz, Martin Kirchmayer, and Leonard Larsen and Arie Poldervaart are preparing papers on the crystalline rocks of the Beartooth Mountains.

Published and unpublished maps and data of the workers named have contributed much knowledge concerning structural detail in specific areas of the Beartooth Mountains, as well as general information concerning Laramide deformation of the Beartooth massif with respect to adjacent areas.

Acknowledgments

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The authors appreciate the use of the facilities of the Yellowstone-Bighorn Research Association at Red Lodge, Montana, of the Departments of Geology at Franklin and Marshall College and Princeton University, and of Stanford Research Institute. The authors take pleasure in acknowledging the able field assistance of Thomas Deery, Merle Logan, Lawrence McKague, Richard Parizek, John Randall, Richard Sheppard, Lawrence Walters, and John Whetton. Each of the writers has profited from discussions in the field and office with John C. Maxwell of Princeton University.

The interest of many Red Lodge citizens in the work of scientists associated with the Y.B.R.A. and the friendliness of the community have contributed to the pleasure of work in this area. Two citizens contributed to this investigation. James Brophy provided information about coal beds and the structure of the sedimentary rocks near Red Lodge. Alan Weaver, with his cheerful companionship, limitless curiosity, and jeep-driving ability, expedited much of the work in physically difficult terrain.

The authors have profited from discussions in the field with Walter Bucher, Willard Parsons, Arie Poldervaart, W. T. Thom, Jr., and other investigators using the Y.B.R.A. Camp. Paul Richards, W. T. Thom, Jr., J. T. Wilson, Jack Fanshawe, and J. Kaspar Arbenz made valuable suggestions in the compilation of the geologic map. The authors particularly appreciate the careful reviews of both text and map by Donald L. Blackstone, Jr., John C. Maxwell, William Pierce, Arie Poldervaart, and John J. Prucha. Their reviews do not necessarily imply their agreement with all the ideas of the authors.

STRATIGRAPHY

Precambrian crystalline rocks constitute the basement complex of the Beartooth Mountains. They include granitic gneisses, amphibolites, minor quartzites, iron formation, mafic dikes, and other irregular masses of mafic and ultramafic rocks. Eckelmann and Poldervaart (1957) and Harris (1959) describe these rock types in detail and interpret them as a thoroughly granitized mass of metasedimentary rocks. Lower-rank Precambrian metasedimentary rocks of Beltian type and age have not been found in the Beartooth Mountains. Along part of the northern border of the Beartooth Mountains, the Stillwater complex constitutes the basement. This complex is the tilted remnant of a differentiated Precambrian lopolith (Hess, 1960) now exposed through a 17,000-foot thickness and a 30-mile length along the mountain front.

Cambrian and younger sediments were deposited upon the crystalline basement complex. Table 1 shows the stratigraphy of these.

The older sedimentary rocks are mainly marine; the Upper Cretaceous sedimentary rocks

are dominantly continental. The Paleozoic rocks total about 3000 feet in thickness. Silurian rocks are missing. Rocks of all other Paleozoic periods are well exposed along the mountain front. The Paleozoic rocks are dominantly carbonates, the Mississippian Madison limestone and the Ordovician Bighorn dolomite being the most easily recognizable units along the mountain front. Structurally less competent units are the Cambrian Gros Ventre and Grove Creek shales, the Devonian Threeforks shale, and the Mississippian-Pennsylvanian Amsden formation.

The Mesozoic systems total about 7000 feet of dominantly fine-grained clastic materials. These rocks, although uplifted along with the

TABLE 1. STRATIGRAPHIC COLUMN, NORTH AND EAST BORDERS OF THE BEARTOOTH UPLIFT, MONTANA

AGE	FORMATION		LITHOLOGY		THICKNESS (Feet)	
	North	East	McLeod Area (North)	Deep Lake Area (East)	North	East
Quaternary			Terrace gravels, floodplain deposits, glacial deposits			
Eocene		Willwood	Absent	Red to tan massive arkoses and conglomerates containing cobbles of sedimentary and Precambrian rocks		2500+
Paleocene	Not mapped	Undivided	Not mapped	Gray to buff bentonitic siltstones and impure sandstones with local thin beds of pebble conglomerate.		
?—?—?—?	Livingston		Volcanic breccias, volcanic-rich sandstones		3500+	2300-2700
	Eagle		Light-gray massive sandstone, gray shale, thin coal beds.		750	
	Telegraph Creek		Light-gray thin-bedded sandstone, interbedded with dark-gray shale.		370	
Cretaceous	Cody		Black shale with thin calcareous concretion zones and thin platy sandstones scattered throughout.		1300	1625
	Frontier		Gray massive to thin-bedded sandstone, dark-gray shale, volcanic-rich (west), thin beds of chert-pebble conglomerate.		630	300
	Thermopolis and Mowry		Dark-gray shale with thin bentonites, thin gray platy sandstones.		975	1165
	Kootenai	Cloverly	Chert-pebble conglomerate (west) or pebbly cross-bedded sandstone (east) at base, red claystone in middle, gray sandstone at top.		450	620
Jurassic	Morrison		Pink and gray shales and claystones, light-gray sandstone.		360	160
	Swift	Sundance	Greenish-gray, highly calcareous glauconitic sandstone.	Pink and tan siltstone and limestone at base, tan siltstone and shale in middle, tan calcareous sandstone at top.	95	610
	Rierdon		Tan and gray calcareous shale, thin oölite bed at base.		160	
	Piper		Pink claystone, gray calcareous shale and platy limestone.		350	
Triassic	Chugwater		Bright-red sandstone, siltstone, and shale		0-110	780
Permian	Phosphoria		Gray vuggy limestone and dolomite, light-gray, very calcareous sandstone.		10-40	40
Pennsylvanian	Quadrant	Tensleep	Light-gray to tan massive sandstone, locally cross-bedded.		40-100	280
?—?—?—?	Amsden		Red shale and siltstone, gray dolomite, gray cherty sandstone and limestone.		80-135	80
Mississippian	Madison		Thin-bedded to massive, light-gray to tan, coarsely crystalline to aphanitic limestone, locally dolomitic, cherty in zones.		1000	780
Devonian	Threeforks		Brown platy dolomite and gray-green shale.	Light-gray to yellow limestone and dolomite, thick calcareous sandstone at base.	70-100	125
	Jefferson		Gray to brown limestone and dolomite, breccias in zones, some fine-grained sandy beds.		345-400	240
		Beartooth Butte	Absent	Local lens of deep-red, medium-bedded, slightly calcareous sandstone.		0-55
Ordovician	Bighorn		Massive, tan, rough-weathering dolomite and dolomitic limestone, thin-bedded brittle dolomite at top.		100-140	310
Cambrian	Grove Creek and Snowy Range		Limestone flat-pebble conglomerate, aphanitic gray limestone, dark-gray shale.		320	420
	Maurice		Gray to brown oölitic and coarsely crystalline limestone, some thin beds of limestone flat-pebble conglomerate.		125	90
	Park		Interbedded dark-gray shale, tan platy sandstone and siltstone, platy aphanitic limestone, and local thin beds of limestone flat-pebble conglomerate.		485	580
	Meagher		Light-gray, wavy-bedded limestone, some interbedded green shale.		100	55
	Wolsey		Dark-gray shale, some platy limestone and siltstone.		40+	210
	Flathead		Light-tan to white, medium-grained sandstone, locally conglomeratic.		0-60	25-60
Precambrian			Green to gray amphibolite and schist, granitic gneiss intruded by basic rocks of the Stillwater igneous complex.	Granitic gneiss, amphibolite and mafic dikes.		

Paleozoic rocks at the mountain fronts, are generally poorly exposed and in large part are covered by debris from the mountains.

The Paleocene Fort Union formation was deposited in the early stages of basin formation. It consists of sandstones, siltstones, shales, and coal. Near Red Lodge, Montana, this formation is about 7600 feet thick. Overlying it are the formations of the early Eocene Wasatch group, locally unconformable and locally containing basement cobbles from the eroded mountains. Uplift and warping of these basin sediments along the mountain front continued through at least part of Wasatch time.

A number of facies changes occur in the Beartooth Mountains (Table 1). Only the most obvious of these are mentioned. The Beartooth Butte formation, laid down as a channel deposit near the shore of the Devonian sea, is found only in the southeast part of the Beartooth Mountains (Dorf, 1934; Pierce, 1957). The Triassic Chugwater formation thins and disappears westward along the range. The thick volcanic rocks of the Livingston formation along the northwest flanks of the Beartooth Mountains result from volcanic activity which began in Judith River (Cretaceous) time. These grade laterally eastward into recognized sedimentary units ranging in age from Judith River to Bearpaw.

From the point of view of the structural development of the massif, the most significant points are:

(1) Approximately 2 miles of Paleozoic and Mesozoic sedimentary rocks covered the basement complex at the beginning of gentle regional folding and block uplift.

(2) Several carbonate formations of the middle and upper Paleozoic section are the most massive and most difficultly deformed of the sedimentary rocks; separating these carbonate formations from the basement is about 1000 feet of shale-rich Cambrian sedimentary rocks which acted as lubricants in deformation and absorbed many of the faults coming out of basement.

(3) Evidence indicates that mild deformation had begun by late Cretaceous times; basin formation had begun in Paleocene time; and major uplift of the block was in full force by early Eocene time.

TECTONIC SETTING

General Statement

The Beartooth Mountains are in the Middle Rocky Mountains, east of the frontal edge of the great belt of overthrusting that extends roughly north-south from northwestern Montana along the eastern border of Idaho into north-central Utah (Fig. 1). Many workers have described the structural features of the overthrust zone: Eardley (1951) gives a good summary.

Uplifts and basins in juxtaposition generally characterize the topographic and structural relationships of the Middle Rocky Mountains. This region is on the Wyoming shelf, which existed as a broad and stable structural platform with relatively shallow marine inundation throughout Paleozoic and Mesozoic time until late Jurassic or early Cretaceous time when more rapid subsidence of the shelf and extensive marine flooding occurred.

The Beartooth is one of the largest mountain masses in the Middle Rocky Mountain region. It exhibits structural features both within its core and along its perimeter that are typical of other uplifts in the region (Bucher, Chamberlin, and Thom, 1934).

Tectonic Pattern of Bounding Areas

Bighorn Basin. Reference should be made to the tectonic map of the region (Fig. 2) in considering the Bighorn Basin and the other structural units.

The Bighorn Basin bounds the Beartooth Mountains on the east and possibly for a short distance along the northeast. The basin is almost completely surrounded by uplifted mountain masses, including the Owl Creek–Bridger massif along the southwest, south, and southeast borders and the Bighorn and Pryor mountains along the eastern border. Only at the north does the basin have no bounding mountain mass. Here the Nye-Bowler zone provides a structural boundary along which the Precambrian basement has moved both horizontally and vertically.

The Precambrian crystalline rocks flooring the basin are covered by as much as 20,000 feet of Paleozoic, Mesozoic, and lower Cenozoic sedimentary rocks and sediments along the basin axis. Andrews, Pierce, and Eargle (1947) describe the axis of the basin as a sinuous line beginning in the southern part of the basin with a northwest trend, then changing to north, to northwest, and again to north close to and along the eastern front of the Beartooth Mountains from Clarks Fork Canyon, Wyoming, to Red Lodge, Montana.

Two features of the axial position of the Bighorn Basin are especially noteworthy. The

is the prominent northwest alignment of the axis between Cody and Powell, Wyoming; the axis strikes directly into the Beartooth Mountain front at the Clarks Fork Canyon and lines up with the prominent Cooke City structural zone that crosses the entire Beartooth massif. The second is that the basin axis as drawn by Andrews, Pierce, and Eargle (1947) along the eastern side of the Beartooth Mountains is so

frontal edge of the Beartooth block at Red Lodge, Montana.

Axes of many smaller folds around the margin of the Bighorn Basin also trend northwest. Along the west side of the basin at the mouths of the Clarks Fork Canyon and Bennett Creek two such folds have been truncated by the bounding fault during uplift of the Beartooth block.

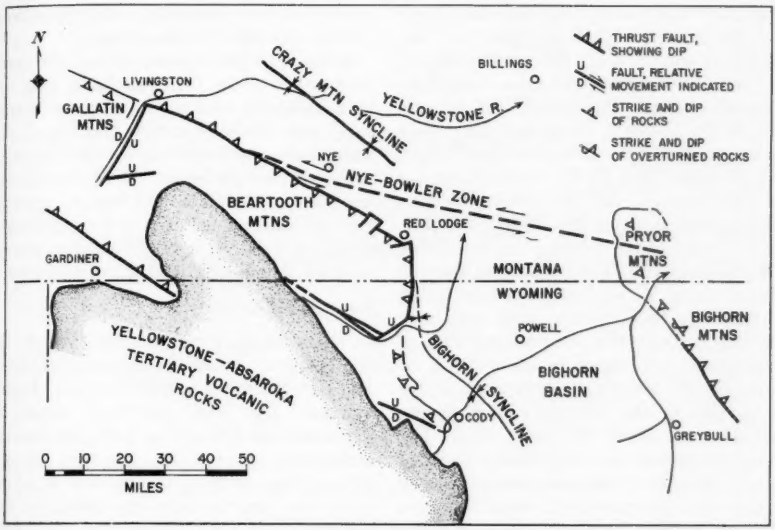

Figure 2. Major tectonic features of the Beartooth Mountains and adjacent areas

close to the mountain front that many geologists since Chamberlin (1940) have speculated that the true axial position is under the faulted margin of the Beartooth block.

Whether their speculation is true or not, there is no record from the extensive coal-mining operations near Red Lodge that any of the workings crossed the basin axis, although none of the mine workings got closer to the mountain front than about a mile (J. Brophy, personal communication). Available seismic information (Buwalda and Gutenberg, 1935; J. R. Fanshawe, personal communication) also indicates sedimentary dips toward the mountain front and an axial position for the basin very close to or under the frontal edge. W. T. Thom, Jr. (Personal communication, map and overlays), conjectures that the Bighorn Basin axis is probably between 8 and 10 miles under the

Alpha and Fanshawe (1954) postulated a narrow extension of the Bighorn Basin to the northwest around the corner of the Beartooth Mountains at Red Lodge. The northern boundary of this narrow prong is the Nye-Bowler structural zone.

Nye-Bowler zone. A group of asymmetric domes and anticlines are almost perfectly aligned about N. 75° W. for nearly 60 miles from the middle of the Pryor Mountains on the east, across the north end of the Bighorn Basin, and along the front of the Beartooth Mountains through the towns of Luther, Roscoe, Dean, and Nye, Montana (Fig. 2). Two fault systems, one parallel to the main zone and one en échelon with N. 50° E. strikes, are associated with the folds. Porphyry laccoliths are concentrated in Paleozoic rocks near the west end of the zone along the Beartooth Mountain front. J. T. Wil-

son (1936, Ph.D. thesis, Princeton Univ.) thought that the structural features associated with this prominent zone are surface expressions of deep-seated movement of basement blocks along a vertical fault parallel to the zone in which the relative movement of the south block was down and to the east. Actually, the sense of movement of the rocks along this zone changes from place to place. A left-lateral strike-slip component of movement exists throughout most of the zone In the vicinity of Red Lodge the south side is down, whereas near Dry Creek and at the eastern end of the zone near the Pryor Mountains the north side is down Movement took place along the zone from late Cretaceous through Paleocene time as evidenced by thickening of the sedimentary rocks south of the zone in the vicinity of Red Lodge,

J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.) has theorized that the Nye-Bowler zone extends westward along the Mill Creek-Stillwater structural zone, dividing the northwest part of the Beartooth Mountains into a separate structural block. Two other possibilities exist: first, the Nye-Bowler zone may terminate against the northwest-trending frontal edge of the Beartooth Mountains in the vicinity of Nye; second, the zone may extend to the northwest, parallel to and incorporated with the frontal structure of the Beartooth massif. For instance, the prominent East Boulder fault has essentially the same trend and comes within 10 miles of direct connection with faults that are a part of the Nye-Bowler zone.

The Nye-Bowler zone separates the northern part of the Bighorn Basin from the Crazy Mountain basin.

Crazy Mountain basin. Bounding the Beartooth Mountains on the north, and west of the termination of the Nye-Bowler zone, is the Crazy Mountain syncline, a northwest-trending structural basin which Thom (1923) considered to have been depressed by the transfer of material from beneath the crust to the surface as volcanic material. Unlike the boundary of the Beartooth Mountains with the Bighorn Basin, where there is sheer structural relief of between 15,000 and 20,000 feet, the structural relief between the Beartooth Mountains and the Crazy Mountain Basin is between 10,000 and 15,000 feet over a distance of 8 to 10 miles. The sedimentary rocks constituting the southwest flank of the syncline rise out of the basin and become a part of the steep, partly faulted flanks of the uplifted Beartooth massif.

Absaroka-Yellowstone volcanic area. Bounding the Beartooth Mountains along almost all of the southwest side are thick Tertiary volcanic rocks consisting primarily of breccias, conglomerates, tuffs, and flows. These obscure the edge of the Beartooth Mountain block and its structural detail (Fig. 2; Pl. 1).

A series of volcanic vents with associated radial dikes occurs along the Cooke City structural zone which constitutes a northwest-trending medial line through the entire Beartooth block. Possibly the vents and associated volcanic rocks are genetically related to and structurally controlled by this zone.

Johnson (1934) speculated on the possible connection of the Gardiner fault, near Gardiner, Montana, with the Rattlesnake Mountain fault, near Cody, Wyoming (*See* Fig. 2), along a continuous structural zone under the volcanic rocks that would bound the southwest side of the Beartooth block. The writers agree with this speculation, but it is as yet unproved because of total lack of knowledge regarding geologic details beneath the volcanic cover.

Geophysical Setting

The available geophysical data on the Beartooth region include fragmentary seismic data and data from a reconnaissance net of gravity stations. The only published seismic data (Buwalda and Gutenberg, 1935) are from three reflection stations on the mountain side of the frontal Beartooth fault zone near Red Lodge, Montana. Results were not definitive, but a suggested interpretation was that a surface dips toward the mountains at about 30°. Ewing and Press (1959) have obtained surface-wave data; they indicate depths to the M-discontinuity of about 47 km in the region of the Middle Rocky Mountains.

Woollard (1936; 1937; 1955) states that the Middle Rocky Mountains, including the Beartooth Mountains, are gravity highs superimposed on a widespread negative gravity area. William Bonini of Princeton University is currently carrying on detailed gravity investigations in the region. A preliminary map of these gravity data exhibited in November 1959 shows a sharp gravity gradient approximately parallel to and along the Nye-Bowler lineament from near Nye to the Pryor Mountains. One interpretation of these data is that there are fundamental differences in the lithology and structure of the deep basement adjacent to this zone. The north flank of the Beartooth block may have been partially defined by these differences.

Woollard suggests that the region as a whole

approaches equilibrium but areas of structural complexity show departures from that equilibrium. He finds that the Beartooth Mountains have a positive gravity anomaly of approximately 80 mgals, whereas the Bighorn Basin has a negative anomaly of about 30 mgals. The anomalies are characterized by sharp breaks, inactivity occurs along these zones, nor are the late Tertiary and Quaternary surficial deposits in these zones disturbed. Therefore, the writers regard the present crustal structure of the Beartooth region as a reasonably stable configuration with regard to the strength of the earth's crust.

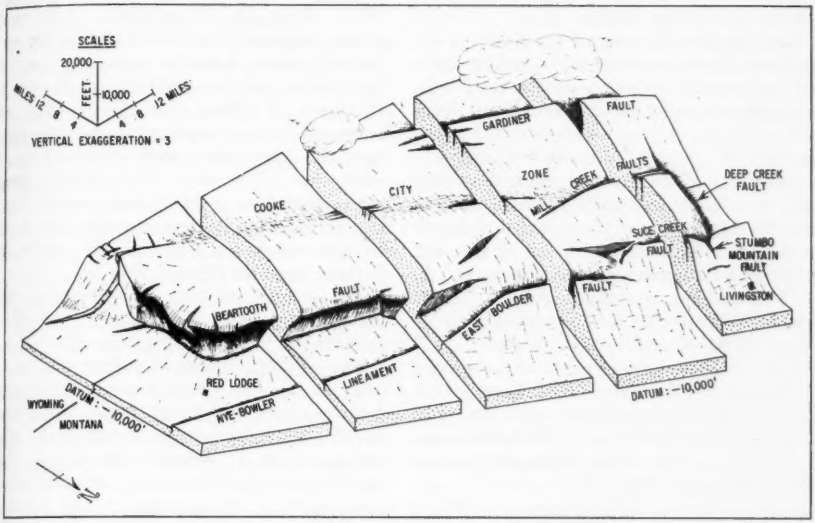

Figure 3. Structural interpretation of the Beartooth block

dicating that at least part of their cause is at shallow crustal depths.

The 110-mgal gravity difference may be explained in two ways in terms of the earth's crust. First, the crustal material immediately beneath the Beartooth block may be denser than that beneath the adjacent basins. Second, the position of the M-discontinuity under the mountain ranges may be higher, thereby bringing the denser, sub-Moho material closer to the surface. Present gravity conditions around the Beartooth Mountains are compatible with either one or a combination of the two situations. However, to explain the gravity data as well as volume differences created by a mountain mass lifted 3–4 miles, one must postulate a rise of the denser, sub-Moho material beneath the mountain blocks.

In spite of the gravity differences of about 100 mgals, the frontal disturbed zones separating the massifs from the adjacent basins appear to be stable at present. No pronounced seismic

STRUCTURE

General Summary

The Beartooth Mountains constitute a massive uplifted crustal block of metamorphosed and granitized Precambrian sedimentary rocks (Eckelmann and Poldervaart, 1957). The core of the massif consists of foliated crystalline rocks transected by several major structural zones and many lesser fractures that divide the block into a few large and many smaller units (Fig. 3; Pl. 1).

The block is topographically and structurally highest along the east and northeast edges. The surface developed upon the exposed crystalline rocks slopes gently southwest toward the Cooke City structural zone where elevations at Precambrian-Cambrian contacts reveal a flattening of the slope. Younger unmetamorphosed Paleozoic and Mesozoic sedimentary rocks are faulted and draped over the steep eastern edge and a part of the northeastern edge, making a striking

array of vertical and steeply dipping rocks expressed as hogbacks and palisades.

Erosional remnants of lower Paleozoic sedimentary rocks, particularly Cambrian, are preserved on top of the Precambrian rocks of the massif (Poldervaart and Bentley map, 1958); they dip southwest, as does the crystalline rock surface. Along the axis of the Cooke City structural zone the sedimentary patches are more extensive (Lovering, 1929).

Along much of the east and north edge of the mountains both Precambrian crystalline rocks and sedimentary rocks overlie younger sedimentary rocks along one or more thrust planes with visible dip toward the mountains ranging between 30° and 90°. The draped rocks are steeply tilted or overturned. Numerous tear faults with displacements of up to 10,000 feet break the upper plate of the thrust to dislocate the mountain front. Along the steep east and north sides the structural relief measured on the Precambrian crystalline rocks is in the order of 10,000–20,000 feet. The attitude of the frontal thrusts at depth is speculative.

Much of the west side of the Beartooth massif is bounded by the Deep Creek fault which separates the high mountain mass from the broad alluvium-filled valley of the Yellowstone River. Displacement along this fault decreases southward toward Gardiner.

In the vicinity of Gardiner there is a reverse fault along which the massif has moved upward, developing a structural relief of at least 12,000 feet (Wilson, 1934). Only at this, the southwest part of the Beartooth Mountains, is there no true block corner. Here the Beartooth and Gallatin ranges are joined together without any structural separation. Tertiary volcanic rocks mask further delineation of the southwestern boundary of the mountain block toward the southeast.

Major Relationships of Basement Structure

Precambrian crystalline rocks. There is a great disparity of knowledge concerning Laramide and Precambrian structures of the Middle Rocky Mountains. In general, geologic mappers have "lumped" the Precambrian rocks under the term "undifferentiated basement complex." Notable exceptions are Lovering and Goddard (1950) in the Front Range, Noble (1950) in the Black Hills, and recent and current workers on the structure and petrogenesis of the core of the Beartooth Mountains, most of whom are part of a large research group from Columbia University directed by A. Poldervaart. The present authors have limited their structural work on Precambrian rocks to the area near the Clarks Fork Canyon at the southeast end of the Beartooth Range.

The Precambrian core of the eastern and southern Beartooth Mountains consists of a series of south- or southwest-plunging folds (Pl. 1) in ancient metasedimentary rocks (Poldervaart and Bentley, 1958; Wise, 1957, Ph.D. thesis, Princeton Univ.; 1958) with granitic material structurally compatible with the folds but of debatable origin. Age dates of this granitic and metasedimentary complex are all about 2.7 billion years (Gast, Kulp, and Long, 1958). Superimposed on the older structures is a Laramide system of closely spaced joints that follow older, fluid inclusion planes in the crystalline rocks. In the southern Beartooth Mountains several Laramide faults follow and reshear a north-northeast-trending set of these fractures (Wise, 1958).

The entire granitic and metasedimentary mass contains many Precambrian mafic dikes. Spencer (1959) reports that these dikes follow a number of Precambrian joint directions. In the southern Beartooth Mountains the dikes have a dominant northwest trend (Pl. 1). In the western Beartooth Mountains several of the dikes mapped by J. T. Wilson (1934, unpub. field map) have a northeast trend. The northwest-trending Precambrian dikes in the southern Beartooth Mountains parallel some of the Laramide faults in that area (Wise, 1957, Ph.D. thesis, Princeton Univ.). The dikes and some joint sets are the only recognized Precambrian structures paralleling the northwest elongation of many of the ranges of the Middle Rocky Mountains. The lines of serpentinized peridotite intrusions through Wyoming (Hess, 1955) suggest a Precambrian alpine orogenic belt of northwest trend. Intrusions, dikes, and deeper crustal structures of this poorly known, poorly established orogenic zone may have controlled the dominant northwest trend of the present ranges and minor structures of the Middle Rockies.

The concept that the central part of the massif consists of a batholithic core seems questionable on the evidence of surface exposures (Poldervaart and Bentley, 1958; Wise, 1957, Ph.D. thesis, Princeton Univ.). Work of Horberg and Fryxell (1942) in the Teton Mountains, Osterwald (1949) in the Bighorn Mountains, and E. W. Spencer (1958, personal communication) in the Madison Range shows strong evidence of ancient metasedimentary rocks as

opposed to a simple batholithic core controlling the mountains. As a working hypothesis the present authors believe that the distribution of the ranges of the Middle Rocky Mountains was defined by Laramide stress patterns, modified by pre-existing northwest-trending structural zones of weakness in the basement. Small-scale features of block shape were determined by the local fracture pattern of the basement and in some cases by lithologic boundaries.

. *Basement faults.* Major structural zones in Precambrian rocks that have or may have played an important role in Laramide deformation include faults that bound the Beartooth block, faults in adjacent structural areas, and faults within the block itself. These types will be discussed separately.

FAULTS BOUNDING THE BLOCK: The massive Beartooth crustal block was uplifted during the Laramide orogeny. The faults or fault zones in the Precambrian rocks that define the boundaries of the block may have been developed during the initial stages of the Laramide orogeny, but it is distinctly possible that they are older fractures, possibly Precambrian. Their alignment in two major directions, about N. 10°–20° E. and N. 60°–70° W., is the same as that of many of the Precambrian mafic dikes that occupy older fractures in the crystalline rocks of the block's interior (Spencer, 1959).

These bounding faults and fault zones along which the block was uplifted are the structures along which the veneer of covering sedimentary rocks draped, folded, broke, and slid from the high frontal edge of the block onto the adjacent structurally lower basins. These faults and associated structures around the entire perimeter of the block are described in detail later.

BASEMENT FAULTS IN ADJACENT AREAS: Only one major structure that cuts the Precambrian basement and along which subsequent movement has deformed the overlying Paleozoic, Mesozoic, and Paleocene sedimentary rocks is near enough to the Beartooth block to be genetically significant. This is the Nye-Bowler zone, which has already been described and which will be further discussed in conjunction with the detailed structural descriptions of the perimeter of the block.

It is noteworthy, however, that two other structural zones in central Montana have the same characteristics as the Nye-Bowler zone. Hancock (1918) described the Lake Basin fault zone. North of it is the Cat Creek fault zone, near the northern part of the Middle Rocky Mountains. The sense of displacement in all three zones is left lateral. Along all three of these zones, fracturing in the overlying rocks and local thickening or thinning of the rocks indicates reactivation of movement along existing structural lines of weakness in the crystalline basement.

Although much of the literature pertaining to the Beartooth Mountains includes the concept of older zones of structural weakness, few facts are known to help establish its validity. Reactivation of movement along pre-existing faults is known, but this may have been due to long-term persistence of the deforming stress rather than to actual weakness of specific crustal zones.

STRUCTURES WITHIN THE BLOCK: Two major structural zones in Precambrian rocks cut across the interior of the Beartooth block. These are the Cooke City zone, described for the first time herein, and the Mill Creek-Stillwater zone, first described by J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.).

The Cooke City zone is a prominent northwest-trending structural zone (Fig. 3) that transects the Beartooth Mountains from terminal points in the Clarks Fork Canyon and the junction of Mill Creek with the Yellowstone River. The features that have been observed along this zone serve to identify it as a major lineament within the Beartooth block.

Fractures are a major feature of the zone. Wise (1957, Ph.D. thesis, Princeton Univ.) and Pierce (1957) have described northwest-trending faults in the Clarks Fork Canyon with vertical displacements of up to 3000 feet. J. H. Sheufler (1954, M.S. thesis, Wayne Univ.) and Spencer (1959) have mapped the fault cutting Beartooth Butte and a host of other prominent fractures in the crystalline rocks having the same northwest alignment in the area near the butte. Precambrian mafic dikes in this area have a dominant northwest trend parallel to the fractures. J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.) described faults in the Mill Creek area with displacements of up to 2000 feet, some having a northwesterly trend. Lovering (1929) and present workers (Pierce, 1957; Parsons, 1958) recognize northwest-trending faults and shear zones in the Precambrian rocks near Cooke City.

The Precambrian crystalline surface slopes southwestward toward the Cooke City zone all along the northwestern part of the Beartooth block. Northwest of Cooke City where it is possible to obtain elevation control at the contact of the Precambrian rocks with overlying

Cambrian sedimentary rocks, the surface reverses southwest of the zone and slopes northeastward, thus outlining a structurally low zone.

Preservation of Cambrian rocks is almost continuous along the Cooke City zone, being interrupted only by several small gaps northwest of Cooke City (Pl. 1).

The incidence of Laramide porphyry intrusive rocks along the zone is greater than in adjacent parts of the Beartooth block. There appears to have been some structural control for at least some of these intrusive rocks; J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.) stated that faulting in the Mill Creek area controlled the position of intrusive rocks there.

Volcanic vents and radial dikes are generally aligned along the Cooke City zone (Pl. 1; Iddings and Weed, 1894; Parsons, 1939; 1958). This suggests that the location of these Tertiary volcanic vents was controlled in part by the prior existence of the structural zone.

Finally, with the exception of the Sunlight district, mineralization in the interior of the Beartooth massif occurs only in three areas, all of which are aligned along the zone. From southeast to northwest they are the Cooke City (Lovering, 1929), Independence, and Emigrant Gulch mining districts (Iddings and Weed, 1894). Some of the mineralization took place after the Tertiary volcanic rocks were extruded. At Independence, for instance, mineralization occurs in the Precambrian crystalline rocks, Laramide porphyries, and the volcanic rocks. The absence of mineralization elsewhere strongly suggests some structural control for the mineralization.

Coincidence of all these features and the interdependence of some on others suggests that this structural zone existed at least since the beginning of uplift of the block and possibly before. As mentioned, the axis of the Bighorn Basin trends at one place almost directly in line with the Cooke City zone, suggesting that the zone may represent the uplifted axis and floor of the Bighorn Basin during an early stage of development. The central, almost axial, position of the zone in the western part of the Beartooth massif may indicate breaking and bowing of the block during its uplift.

J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.) described east-west and northwest faulting in the vicinity of Mill Creek in the Mill Creek-Stillwater zone at the west edge of the Beartooth block. Along these faults, several large patches of Paleozoic rocks as young as the Madison limestone are preserved in downfaulted blocks on the south side of the faults. Ten miles east of the eastern terminus of the Mill Creek fault, along Fourmile Creek, is another patch of rocks from Cambrian to Jurassic in age preserved in a downdropped block on the north side of an east-west fault (Pl. 1).

Primarily on the basis of these faults which have the same alignment and because of the suggestive alignment of drainage and the presence of shearing in some of the intervening Precambrian rocks, J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.) extended the zone 12 miles to the Beartooth Mountain front along the Stillwater drainage. This east-west zone about 35 miles long outlines a large triangular mass of the northwest part of the Beartooth block, which Wilson called the North Snowy block. The Tectonic Map of the United States (King, 1944) shows this zone as a continuous fault.

J. T. Wilson (1936, Ph.D. thesis, Princeton Univ.) has also proposed a possible connection of this zone with the Nye-Bowler zone with a major change of strike at the mountain front. The writers think that there is no tangible evidence to support this and have discussed elsewhere other possibilities of extension of the Nye-Bowler zone.

Work by Howland (1955) and R. Butler has caused Butler (1959, personal communication) to suggest another interpretation of the eastern termination of the Mill Creek zone. He finds that a major discontinuity between metasedimentary material on the north and dominantly granitic material on the south persists along the West Stillwater River; that the West Stillwater River marks a sharp change between overturned dips in the Stillwater complex on the east and moderate normal dips on the west; that the western limit of major frontal thrusting occurs at the West Stillwater River; that the base of the Stillwater complex shows an apparent left-lateral displacement of 1–2 miles along the West Stillwater River (Pl. 1); and that a large thrust (the Bluebird fault of Howland, 1955) cuts out the lower members of the Stillwater complex before disappearing beneath the alluvium of the West Stillwater River. For these reasons Butler has suggested that Wilson's Mill Creek zone may continue eastward along the West Stillwater River to merge with the Bluebird fault. The left-lateral displacement on the zone would have been absorbed near the mountain front by thrusting of basement gneiss over the lower Stillwater complex.

In the two areas where sedimentary rocks are preserved farther west along the zone, maximum vertical displacement along the downdropped blocks probably does not exceed 3000 feet. At the west end of the zone, folds in the rocks plunge steeply (30–50°) toward the Yellowstone Valley. This may be due to downward drag along the Deep Creek fault that bounds the western edge of the Beartooth block, or the folds might have been tilted westward by Tertiary intrusive rocks.

It should be noted that the western end of the Mill Creek zone is the same as the northwestern end of the Cooke City zone. Earlier suggestions that Tertiary intrusions have followed the Mill Creek zone seem to be incorrect inasmuch as none have been mapped east of the Mill Creek terminus. Intrusive rocks have been emplaced at Cooke City and at Independence along the Cooke City zone; the writers think that the influence of the Cooke City zone accounts for their presence at Mill Creek.

Sedimentary veneer. As previously noted, the ranges of the Middle Rocky Mountains are generally considered to result from active movement of large basement blocks, the sedimentary-rock veneer having played a relatively passive role in deformation (Chamberlin, 1945). This is best illustrated at the Buffalo Bill Dam in the Rattlesnake Mountain structure west of Cody, Wyoming. Here a major fault tilting the basement block of Rattlesnake Mountain passes upward into the sedimentary-rock veneer. Much of the displacement of the fault is absorbed in the Cambrian shales. Studies of closely spaced joints (and other fractures) in the basement by the authors show that the crystalline rocks have not been rotated or internally deformed during the development of the fold. The fault displacement gives way in the upper Paleozoic rocks to a relatively simple asymmetric anticline draped across the broken basement. Were it not for the deep exposure in the canyon, one might interpret the asymmetric fold in a number of ways without realizing the basement control of the structure.

The peripheral structures of the Beartooth massif present similar problems. They are interesting as structures alone but their real significance is as indicators of movements and stress patterns of the great crustal block. The purpose in describing these peripheral structures is to reveal the nature of movement along the major faults in the basement directly controlling the movement of the crustal block. Knowing these movements, and their relationship to each other, one can speculate on the fundamental mechanisms involved in the rise of the crustal block.

Perimeter of the Beartooth Block

Beartooth fault zone. Total vertical displacements across the Beartooth block edges are as much as 20,000 feet. Several types of structures and kinds of mechanisms are necessary to achieve the separation. The complex of frontal structures along the east and northeast sides of the Beartooth block is termed the Beartooth fault zone, first noted by Eldridge (1894); along the western side, the Deep Creek fault zone in the Yellowstone Valley; and along the southwest edges of the block, the Gardiner fault zone. In the following discussions the physical characteristics of the zone will be described from Wyoming northward to Red Lodge, Montana, northwestward to Livingston, Montana, south along the Yellowstone Valley to Gardiner, Montana, and thence southeastward to its disappearance beneath volcanic cover. Reference to Figure 3 and Plate 1 will aid the reader in following the discussion.

South end of the Beartooth fault zone. The Beartooth fault zone separates the Beartooth massif from the Bighorn Basin as far south as the Clarks Fork Canyon, Wyoming, where it meets the Cooke City structural zone (Fig. 4). Southward the structural distinction between the Bighorn Basin on the east and uplifted block on the west continues without the presence of major basement faulting. Instead the basin floor rises in a broad ramp onto the gently dipping (10–20-degree) slopes of Dead Indian Hill (Fig. 5). The total rise of basement on this slope is about 20,000 feet to an elevation of 6000 feet above sea level at the top of the ramp.

Near its southern termination the Beartooth fault zone decreases in throw as the vertical displacement is taken up on the Dead Indian ramp. Where last seen before its disappearance beneath the alluvium of the Clarks Fork Canyon the fault dips toward the basin at about 85° and has a vertical displacement of about 3000 feet (Fig. 5; Pl. 2).

The termination of the Beartooth fault zone is related to its intersection with the Cooke City structural zone. Trending southeast in the Clarks Fork valley and directly aligned with the Cooke City zone is the Clarks Fork fault, the Dillworth fault of Hughes (1933; 1932, Ph.D. thesis, Johns Hopkins Univ.), a vertical fault with a 3000-foot displacement separating the basement surface in the Clarks Fork valley from

that of the uplift to the north (Pl. 2, fig. 2). Like the Beartooth fault, the Clarks Fork fault disappears beneath the alluvium of the Clarks Fork Canyon. Neither fault appears in almost perfect exposures on the opposite side of the canyon. The two faults must join at the canyon to form a simple vertically raised corner of the

nor in the corner formed by the vertical fault is there evidence of strike-slip movement or horizontal displacement. *This corner firmly* the uplifted block to the Bighorn Basin and offe some control on the amount of horizontal di placement possible between the massif and t basin in other areas.

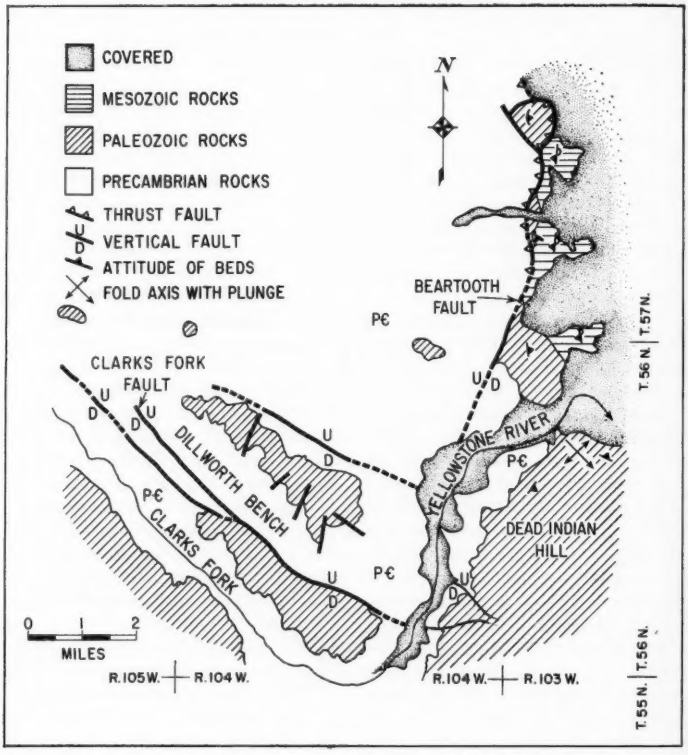

Figure 4. Major structural features at the south end of the Beartooth Mountains

range (Fig. 5). The meeting of these two zones at nearly 90° produced a more efficient break than would have been achieved by an acute angle. A subsidiary fault angles out of the Clarks Fork fault zone to make the right angle juncture and to leave a large, slightly less uplifted fault slice, called by some the Dillworth bench (Figs. 4, 5).

The outstanding characteristic of the structural features of this corner of the range is their domination by vertical movement. Neither in the basement structure of Dead Indian ramp

In the area north of Clarks Fork Canyon, th Beartooth fault can be observed to becom steeper with structural depth. The dip of th fault can be obtained in three areas at three di ferent structural depths. At the Clarks For Canyon, the structurally deepest position o both walls in basement, the fault is nearly ver tical and is parallel to prominent closely space Precambrian joints which apparently contr its position. In Littlerock Creek, 4 miles nort of the Clarks Fork Canyon, at an intermedia structural position of basement rocks in th

hanging wall against sedimentary rocks in the footwall, the fault dips toward the mountains at 50° in the valley bottom. The fault flattens upward within 1000 feet to where exposures show only a 20-degree dip. At Bennett Creek, 7 miles north of the Clarks Fork Canyon, the fault is an almost horizontal thrust. This is the structurally highest position with sedimentary rocks in both the hanging wall and the footwall. The thrust places nearly vertical Cambrian

In two areas, where Bennett Creek and the Clarks Fork of the Yellowstone River exit from the mountains, the Beartooth fault zone cuts and uplifts fold structures similar in type and trend to those of the Bighorn Basin. Both structures are northwest-trending drape folds associated with nearly vertical basement faults along which the northeast side has dropped several thousand feet (Figs. 4, 5). If these folds are truly structures associated with formation

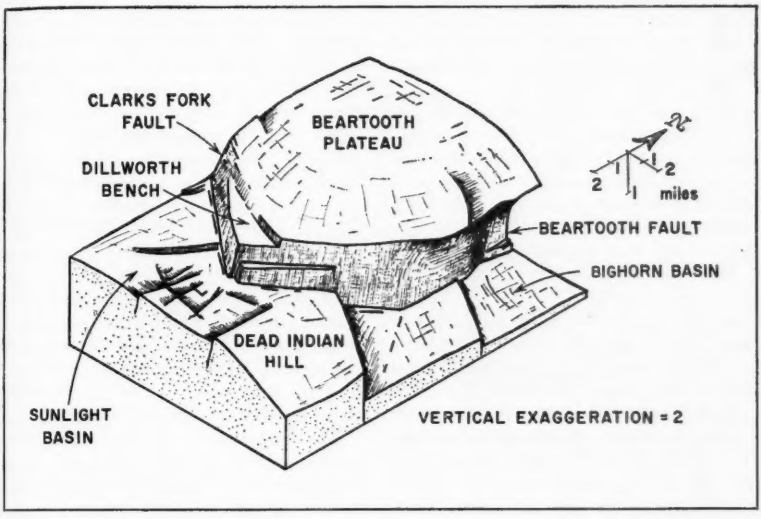

Figure 5. Inferred configuration of the basement at the Clarks Fork corner (Deep Lake quadrangle) of the Beartooth Mountains, Wyoming

rocks on nearly vertical Jurassic rocks, a horizontal displacement of about 1 mile.

The change in dip with depth of frontal faults is a fundamental problem of structural interpretation in the Middle Rocky Mountains. If the dip steepens with depth, horizontal movement is minimized; if it lessens, horizontal movement is increased. Recent model experiments by Sanford (1959) provide a theoretical basis for the steepening of thrust fault dips with increased depth.

In summary, near the south end the Beartooth fault is a thrust with a maximum horizontal displacement of 1 mile, steepening downward into a vertical fault in the basement rocks. At its southern termination it joins the vertical Clarks Fork fault to make a right angle at the uplifted block corner.

of the Bighorn Basin, then major block uplift slightly postdates basin formation, and the block might be expected to truncate basin fold axes. The main basin axis trends northwest, turning sharply northward along the Beartooth Mountain front. The projection of this northwest trend is on a direct line with the Cooke City zone beyond the mountain front. Thus the possibility must be considered that the Cooke City zone is controlled by the uplifted axial extension of an *early developmental stage* of the Bighorn Basin.

Line Creek structure. At Line Creek on the Wyoming-Montana State line, the Beartooth fault zone is complicated by folding of the upturned sedimentary rocks along a vertical axis (Pl. 3, fig. 1). The upper Paleozoic formations are sharply folded but unbroken. The lower

Paleozoic formations are broken, compressed, and shortened by fault overlap. The Bighorn dolomite shows one-quarter of a mile of shortening by fault overlap (Pl. 1).

One might interpret the Line Creek fold upon casual inspection as a drag around a basement tear fault. However, the lack of basement displacement at the front and the frontal shortening by fault overlap invalidate such an interpretation. On the basis of the straight trace of the basement, the faulted overlap of the sedimentary rocks, and the vertical fold axes, the authors interpret the Line Creek fold as a great drag structure caused by northward movement of the mountain block with respect to the uplifted frontal structures.

Two miles south of Line Creek a small tear fault across the Beartooth fault zone is in turn displaced by secondary faults. Like the Line Creek structure these secondary faults indicate a northward movement of the mountain block after most of the uplift was completed. The tear fault is not well exposed, but the Line Creek structure supports the interpretation offered.

The tectonic implication of both these structures near Line Creek is that they are modifications of the frontal structures and hence must be slightly younger than the major uplift of the massif. Both suggest right-lateral displacement. In the writers' opinion, during the later stages of uplift the Beartooth Mountain block began to show lateral displacement toward the northeast with respect to the adjacent basin. This is toward the northeast corner of the block at Red Lodge. The fault overlap and fold shortening indicate that the strike displacement is at least one-quarter of a mile and probably does not exceed 1 mile.

Line Creek to Gold Creek. North of Line Creek on the Montana-Wyoming line, the main Beartooth fault trace passes out through the Paleozoic sedimentary rocks and is hidden under the alluvium accumulated along the mountain front. Some patches of Fort Union and Wasatch rocks near the front provide the only spatial control for location of the main fault (Pl. 1).

West of the main fault are three high-angle faults. One of these places Precambrian rocks against Cambrian and Ordovician; the other two create imbricate slices in the overturned sedimentary rocks. Exact dips cannot be measured on these faults, but field relationships suggest that they are steeper than the overturned Paleozoic sedimentary rocks which dip 50° toward the mountain. This is the first part of the mountain front in which imbrication is a significant structural feature. Degree of imbrication increases toward the northeast corner of the block at Red Lodge.

Gold Creek to Harney Creek. Between Gold Creek and Harney Creek the trend of the frontal edge of the Beartooth block changes sharply from N. 15° E. to N. 60° W. The sharp bend constitutes the northeast corner of the uplifted block near Red Lodge, Montana (Fig. 6).

This corner of the block is totally different from the corner at the Clarks Fork of the Yellowstone, where two vertical faults physically tie the Beartooth block to the adjacent Bighorn Basin, not allowing for any horizontal movement. Here three prominent structural features provided a mechanism for displacement of the frontal edge of the block outward over the adjacent sedimentary-rock-filled basin (Fig. 6; Pl. 1). These features, described by Foose (1960), are the major fault itself, a series of smaller thrusts in the upper plate of the main fault block that break it into imbricate slices, and numerous tear faults that break across the upper plate and allowed large masses of the upper plate to move outward as much as 2 miles. Only in this corner of the Beartooth block is there significant imbrication and tear faulting.

The major fault is not visible anywhere around this section of the mountain front. The location of Fort Union sedimentary rocks close to the mountain front requires a fault to account for missing stratigraphic thickness and provides an approximate location of the fault beneath alluvial material. Probably the thrust places lower Cretaceous rocks onto the Fort Union. The attitude of the main fault is not known and may be inferred only by one's assuming that it has a dip near the surface similar to one or more of the smaller imbricate faults above it. On this basis, its dip toward the mountain would be between 30° and 45°.

Imbrication of the Paleozoic sedimentary rocks along this section of the front is marked. Between Rock Creek and West Rock Creek there are three prominent imbrications above the main fault. The Paleozoic limestones are vertical or steeply overturned and crop out in the most striking hogbacks and palisades in the entire mountain uplift. The several thrusts have caused both elimination and repetition of these formations. Throughout this section of the front the uppermost fault causes the Precambrian crystalline rocks to lie upon Cambrian

Ordovician, Devonian, and Mississippian formations. Adjacent to Rock Creek this fault dips toward the mountain at 35°.

Three large tear faults break and offset the mountain front near the northeast corner (Fig. 5; Pl. 1). At Red Lodge the Maurice tear fault has left-lateral displacement of 10,000 feet (Pl. 2, fig. 2). Northwest of Red Lodge the Willow Creek tear fault has right-lateral displacement in the order of 3000 feet, and the Harney Creek tear fault has left-lateral displacement in the order of 4000 feet. These tear faults break the mountain front into major blocks; they allowed parts of the bold frontal edge to move considerable distances over the sedimentary-rock-filled basin.

There has been much speculation concerning the amount of horizontal displacement of the Beartooth block over the Bighorn Basin (Fig. 1). The writers believe that the maximum horizontal displacement is at the northeast corner of the block. Both the Clarks Fork and Livingston corners of the range are tightly fixed to the adjacent basins as shown by their structural relationships. All the deformational stresses culminated at the Red Lodge corner of the uplift and served to break and move forward the frontal edge. Along the Maurice tear fault the minimum apparent displacement is 10,000 feet.

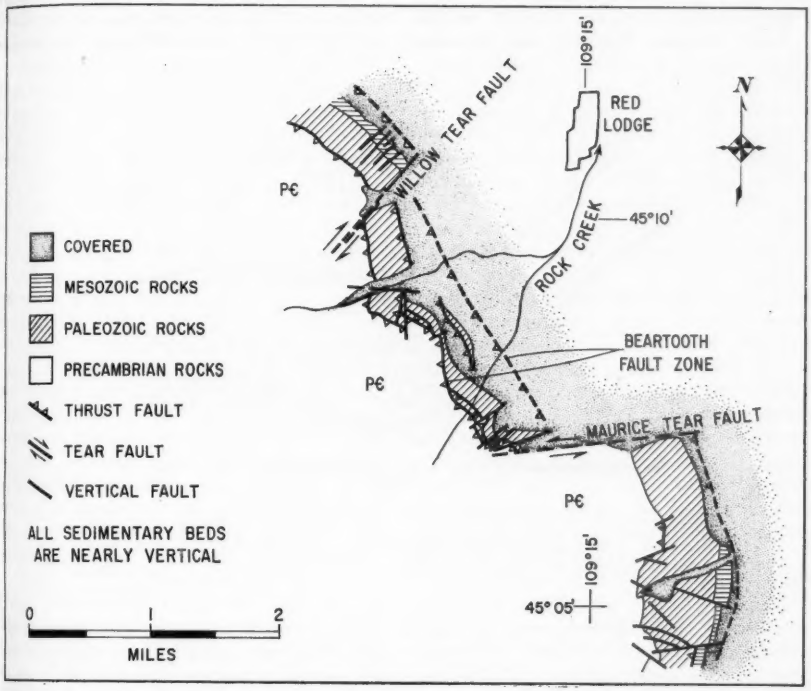

Figure 6. Major structural features of the Beartooth Mountains near Red Lodge, Montana

Some additional horizontal movement may have taken place as a result of the imbrication and by movement along the major fault. Recognizing that the major fault is vertical at the Clarks Fork corner, only 20 miles south of the Maurice tear fault, the authors believe that the attitude of the Beartooth fault at the Red Lodge corner becomes vertical or very steep at depth. Until the attitude of the fault at depth is actually determined, this and other statements regarding the total amount of horizontal displacement are speculative.

Harney Creek to East Rosebud Creek. This 8-mile-long section of the mountain front trends N. 60°–70° W. It is the first section of the

frontal edge northwest of the greatly faulted corner at Red Lodge. The absence of tear faults is noteworthy. Complexly intermingling thrust and lateral shear structures characterize this section. These structures are particularly well displayed between Burnt Fork and Red Lodge Creek, midway in this part of the mountain front (Fig. 7). Three major and several minor faults, all having left-lateral displacement, have served to crumple, fold, and shear the vertical or steeply dipping Paleozoic and Mesozoic

location of Wasatch and Fort Union sedimentary rocks close to the mountain spatially control its position, close to the outcropping Sundance rocks.

East Rosebud Creek to Fishtail Creek. For 14 miles the frontal edge of the mountain is boldly elevated mass of Precambrian crystalline rocks with no exposed sedimentary rocks (Pl. 1). Glacial moraine and outwash and erosional debris from the mountain conceal any Paleozoic or Mesozoic rocks that might be present

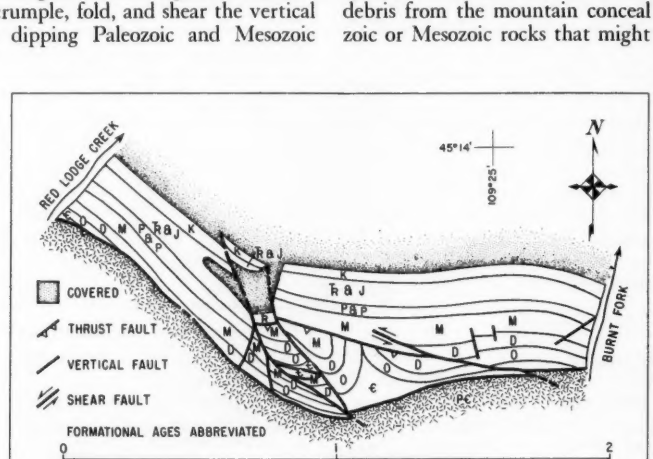

Figure 7. Complex thrust and shear structure near Red Lodge Creek, Beartooth Mountains, Montana

rocks which had already been imbricated by a series of high-angle reverse faults. Shortening of the lower Paleozoic section by this shearing and crumpling amounted to a minimum of half a mile and possibly as much as 1½ miles. The writers believe that these effects have been caused by movement of the mountain block eastward with respect to the uplifted and imbricated frontal structures in a fashion similar to that in the Line Creek area (Pl. 3, fig. 1).

West of the laterally sheared and imbricated structures the exposed sedimentary rocks are broken only by a subsidiary thrust fault that slices away the Paleozoic rocks, placing Precambrian crystalline rocks on Cambrian, Ordovician, Devonian, and Mississippian formations. At East Rosebud Creek, the trace of this fault and all the exposed sedimentary rocks disappear beneath glacial debris.

The main Beartooth fault is not visible at any place along this section of mountain. Again, the although possibly the crystalline rocks have moved outward along the main Beartooth fault far enough to completely cover the more resistant Paleozoic and lower Mesozoic rocks.

West Fishtail Creek to the Stillwater drainage. The next appearance of sedimentary rocks westward along the Beartooth Mountain front is in the vicinity of Fishtail Creek. John S. Vhay (1934, Ph.D. thesis, Princeton Univ.) did the basic geologic mapping in this area. The Westructural Fishtail area has several features of real or potential structural significance. The mountain front at this point shows a discordance in strike. The discordance is reflected in the frontal sedimentary rocks west of the creek where they turn from a northwest to a north-south strike. On the east side of the creek is a small outcrop of Cambrian sedimentary rock and the underlying Stillwater complex. These rocks might be extended along strike to join with the base of the Stillwater complex and some of the

faulted Cambrian sedimentary rocks that are on the west side of West Fishtail Creek. Faulted blocks of Cambrian sedimentary rocks, tilted toward the north where they abut against basement faults, are characteristic of this part of the range front.

The authors do not recognize a unique explanation of the interrelationships of the above facts on the Fishtail structure.

The best explanation seems to be that the location of the initial faults bounding the massif was controlled by the transection of the Stillwater complex, with later fault modification. At the west end of the Stillwater complex near the Main Boulder River an analogous discontinuity of range front appears which results in large part from faulting (Richards, 1957). In like fashion the numerous faults indicated by the indropped Cambrian sedimentary rocks in the Fishtail area may have permitted lesser uplift and consequent folding of the mountain front as they meet the Beartooth fault zone to the east.

Faulting of the basement rocks against the upturned frontal sedimentary rocks continues from Fishtail to the vicinity of the West Fork of the Stillwater River. The frontal faults are slightly en echelon and trend more westerly than the mountain front itself, thus cutting down-section toward the west. Locally, these faults place basement rocks against sedimentary rocks as young as the Madison limestone. One may speculate that these faults represent the reappearance of the Beartooth fault zone from the east. Both the en echelon character and the lessening total displacement of the faults seem to be related to the changing character of the Beartooth front, from thrusting in the east to relatively simple tilting in the west.

West Stillwater River to West Boulder River. The structure of the 20-mile portion of the range front between the West Stillwater and the West Boulder rivers differs from that along the Beartooth fault zone farther east. The structural relief between the Crazy Mountain basin and the Beartooth massif is not as great as in the eastern Beartooth Mountains, and it is spread over a greater horizontal distance (Fig.). A number of faults and tilted fault blocks are associated with this more gentle rise of the basement toward the south. Largest of these is the East Boulder fault.

The steeply dipping faults in the basement, commonly preserving Cambrian sedimentary rocks on their downthrown south side, have the opposite sense of displacement from the uplift of the main mountain mass on the south. A number of folds in the sedimentary rocks are parallel to the mountain front along its northern edge. Associated with these folds are a number of subparallel faults which have the same south-side-down sense of displacement. The folds may be drape structures across similar basement fault blocks, but this cannot be proved by the surface mapping. Where fault dips can be seen they are steep; at no place have large thrusts been mapped.

Some authors (Lammers, 1937; D. C. Skeels, 1939, Ph.D. thesis, Princeton Univ.), thinking that thrusting must inevitably be associated with block uplift, have discussed these structures as thrusts and considered the problem of overthrusting versus underthrusting. In view of the steep dips observed at the surface, the present authors think that the term "thrust" has misleading genetic connotations and that the term "reverse fault" should be applied to those structures with steep north dips. These north-dipping reverse faults are downthrown on the range side and result from rotational movement of blocks that have tilted in the opposite direction to the broad ramplike uplift of the northwest Beartooth front. The authors think that the rotated blocks in the northwest Beartooth Mountains are related to Laramide deformation and are similar in age to rotated block structures overlain by undeformed Tertiary sedimentary rocks in the Badwater area, Wyoming (Tourtelot, 1953).

The proximity of the Nye-Bowler lineament complicates the structure of this portion of the mountain front. A number of laccolithic intrusions near the juncture of the Nye-Bowler zone with the mountain front probably result from easier passage through the fractured area. Rouse *et al.* (1937) suggest that some of the domal structures in this area, particularly Limestone Butte on the lower reaches of the West Stillwater River, are the result of similar laccoliths at depth. Seven miles west of the area disrupted by the laccoliths and in almost a direct line with the Nye-Bowler fault zone, the East Boulder fault may be observed extending westward for 15 miles from the laccolith zone. The authors believe that these two faults represent parts of the same zone passing through the laccolithic complex and that the East Boulder fault is a part of the frontal structure of the Beartooth Mountains (Pl. 1). The authors believe that the Nye-Bowler zone joins with and is incorporated into the frontal structure of the western part of the Beartooth massif. Alterna-

tive possibilities are treated with the discussion of the Nye-Bowler zone.

Livingston corner. The Livingston corner of the massif is formed by the intersection of northeast-trending structures of the Yellowstone Valley with the frontal structures of the Beartooth uplift. Basic geologic mapping in the area is by Lammers (1937), D. C. Skeels (1939, Ph.D. thesis, Princeton Univ.), and Richards (1957). Similar frontal structures of the sedimentary rocks continue westward into the

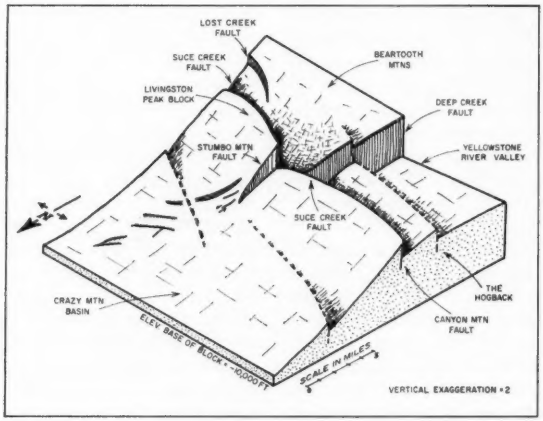

Figure 8. Inferred configuration of the basement in the Livingston corner of the Beartooth Range, Montana

Gallatin Range. As much as 15,000 feet of structural relief separates basement rocks in the Gallatin Range from those in the Beartooth across the Yellowstone Valley.

The Livingston corner area consists of a double corner for the Beartooth Range: the true corner and the Livingston Peak block (Fig. 8). The Livingston Peak block is structurally almost identical to the true block corner, but the Suce Creek fault separates the two. Both the true corner and the Livingston Peak block are tilted approximately 45° northward. The Deep Creek and the Stumbo Mountain faults truncated the western ends of the respective tilted corner blocks and permitted their rotation northward.

Because of the similarity in strike and general features of the Deep Creek and Stumbo Mountain faults, Richards (1957) stated that the two are probably the same fault offset by right-lateral displacement along the Suce Creek fault. The authors do not concur because of (1) the close association of the individual fau with tilting of the corner blocks; (2) lack lateral displacement on the western projectio of the fault in the Canyon Mountain area; a (3) lack of clear evidence of offset at the inte section.

Richards (1957) noted that the Deep Cre fault extends southward beneath the alluviu along the east side of the Yellowstone Valle Outcrops of Madison limestone along the e side of the valley floor, approximately 1 mi

lower than the level of basement rocks im mediately to the east, substantiate this inte pretation. Considerable structural relief persi southward at least as far as the Mill Creek are but here the alluvial cover obscures exact r lationships.

Hence, the Deep Creek-Stumbo Mountai fault zone is significant because it sharp bounds the Beartooth massif along the Yello stone Valley. The northwest-plunging folds this corner of the Beartooth massif also co tributed to the lowering of the structural elev tion of the block toward the Livingston corne These folds continue beyond the Deep Cre Stumbo Mountain fault zone northwestwa toward the Bridger Range.

The Suce Creek fault, separating the Livin ston Peak block from the main block corner, a high-angle reverse fault with approximate 3000 feet of vertical displacement (Richard 1957) near its eastern terminus (Fig. 8). We ward across the Stumbo Mountain-Deep Cre

faults, the vertical displacement of the Suce Creek fault increases to about 5000 feet. Westward beyond the Yellowstone River, the Canyon Mountain structure of D. C. Skeels (1939, Ph.D. thesis, Princeton Univ.) lies on strike with the Suce Creek fault, and basement rocks are exposed along its uplifted north side. Skeels interpreted the structure as a very complex thrust mass pushed southward onto an erosional surface. West of this complex the fault is again nearly vertical to its termination. In view of the broad tectonic setting, the structural detail of adjacent areas, the nearly vertical nature of the fault zone on both sides of the Canyon Mountain structure, and the strange structural complexity in an area of extensive landslide topography, the present authors suggest that the structural complex at Canyon Mountain may be a simple, nearly vertical basement fault modified at the surface by gravity collapse of the overlying sedimentary rocks. Hence they do not regard the complex apparent thrust masses southwest of and below the crest of Canyon Mountain as having special significance in the interpretation of Laramide tectonic forces. Gravity features like this are common in Wyoming, associated with both low- and high-angle reverse faults (D. L. Blackstone, Jr., 1959, personal communication).

Northward from the Livingston corner there are several northeast-trending faults that have apparent left-lateral strike-slip movement as they cut across the sedimentary rocks of the mountain front and adjacent basin anticlines (Richards, 1957). The sense of displacement is such as to have permitted the block corner to move northward with respect to the Yellowstone Valley and Gallatin Range to the west. Along the entire zone cut by these faults the total apparent lateral displacement of the line of outcrop of sedimentary units is only 1 mile. Part of the apparent displacement results from erosional migration down the dip of sedimentary rocks on the uptilted block. A maximum lateral displacement of half a mile seems to be a reasonable estimate.

The tectonic significance of the Livingston corner of the massif is similar to that of the south end of the massif. Both corners are dominated by vertical movement (Figs. 5, 8). The south corner has almost no horizontal displacement with respect to the basins; the Livingston corner has moved less than 1 mile horizontally, probably about half a mile. These two corners, raised vertically with respect to the adjacent basins, provide the fundamental control for setting limits on the total amount of thrusting possible at other places along the edges of the mountain block.

Gardiner area. The Gardiner area, Montana, represents the only certain exposure of the structural relationships along the southwest side of the Beartooth massif. The major feature is a northwest-trending fault along which basement rocks are exposed on the northeast side at least 10,000 feet higher than the sedimentary rocks on the south side (Fig. 2; Pl. 1). C. W. Wilson, Jr. (1934), mapped the geology of this area along the 13-mile exposed length of the structure. Wilson, like Johnson (1934), speculated that the fault or fault zone may project approximately 60 miles beneath the volcanic cover to join with the basement fault of the Rattlesnake Mountain anticline near Cody, Wyoming.

Wilson notes that the Gardiner thrust dips northward at 35° and that several imbricate slices of sedimentary rocks are south of the fault. The fault trace is irregular, presumably as a result of associated tear faulting. The fault trace passes beneath volcanic cover at both ends and, hence, must predate the volcanic extrusion. In Pleistocene or Recent time, the older fault localized basaltic lavas and hot springs deposits (Wilson, 1934).

The exact relationship of the Beartooth block near Gardiner to the adjacent masses of basement rock is not clear because of the thick volcanic cover. Westward, on strike with the Beartooth uplift, the basement rocks in the Gallatin Mountains are at about the same elevation as those in the Beartooth Mountains. This would indicate that the great structural relief produced by the Deep Creek fault in the lower Yellowstone Valley does not persist as far south as Gardiner. Speculatively, the junction of the Deep Creek fault with the Mill Creek or Elbow Creek faults beneath the alluvial cover may permit the southern termination of the Deep Creek fault zone. From a structural viewpoint, there is little or no distinction between the Beartooth and Gallatin ranges in the vicinity of Gardiner. Here the two ranges are apparently joined together without any structural separation. Actually, the Gardiner fault provides a common southern boundary for part or both the Beartooth and Gallatin ranges.

GEOLOGIC CHRONOLOGY

Precambrian

The geologic chronology of the Beartooth region can be subdivided into an ancient sequence of events which produced the basement

rocks constituting the core of the range and a later sequence of events which resulted in the creation of the modern range.

The general chronology of events in the Precambrian core of the Beartooth Mountains is not unique to the Beartooth. A large area including much of the Wyoming shelf has a similar history. Knowledge of the detailed geologic chronology of the Beartooth core is the result of recent and current work, largely by a research group from Columbia University under the direction of Arie Poldervaart (Eckelmann and Poldervaart, 1957; Poldervaart and Bentley, 1958). The following summary is taken largely from their works.

In early Precambrian time a thick sequence of rocks, including quartzites, ironstones, and probably argillites, marls, and subgraywackes was deposited in the region. These were folded along axes which now plunge south or southwest at moderate angles in the southern and eastern Beartooth Mountains. A number of ultramafic bodies were injected into the area in a regional pattern which Hess (1955) suggests is the result of a northwest-trending alpine belt. This was followed by a period of intensive granitization, about 2.7 billion years ago, which produced the complex of amphibolites, biotite gneiss, migmatites, and leucogranitic gneiss, which form the bulk of the Beartooth area. Prior to, synchronous with, and closely following the granitization, mafic igneous activity was widespread. Mafic dikes of the later stages of this activity were emplaced along a number of Precambrian fracture directions (Spencer, 1959). The intrusion of the Stillwater complex may also be associated with later stages of the mafic activity. A number of unmetamorphosed younger Precambrian quartz dolerite dikes cut all other Precambrian rocks.

Through later Precambrian times other sediments may have been deposited in the area. If so, these were removed by erosion prior to the beginning of Cambrian deposition.

Post-Precambrian

Beginning in middle Cambrian time, marine inundation of the eroded surface of metasedimentary crystalline basement throughout the area began. (Refer to Table 1.) The sea was shallow because this area was part of the broad Wyoming shelf, a stable platform that covered much of the Middle Rocky Mountain region. Deposition throughout Paleozoic time was nearly continuous, punctuated by only one substantial epeirogenic uplift during Silurian and early Devonian time and by other smaller uplifts at later times.

During the Triassic, Jurassic, and early Cretaceous, deposition within shallow seas on the stable shelf continued. Thickening of deposits during the late Cretaceous marked the beginning of more rapid shelf subsidence and initiation of early Laramide crustal instability. Part of this instability was manifested by the beginning of volcanic activity along the northeast border of the present Beartooth block and the northwestern extension of the Nye-Bowler lineament (Rouse, 1937). Livingston volcanic rocks are interbedded with rocks as old as the Upper Cretaceous Judith River formation (Parsons, 1942). Doming and possibly incipient faulting occurred at the same time along the Nye-Bowler zone and probably along much of the perimeter of the Beartooth block, particularly the east and northeast sides. Thickening of upper Cretaceous sedimentary rocks on the south side of the Nye-Bowler zone near Red Lodge with respect to the north side indicates that that portion of the southern block moved downward during sedimentation.

Major uplift of the Beartooth block occurred during middle and late Fort Union (Paleocene) time and culminated during early Wasatch (Eocene) time. The early upward movement of the block was along bounding faults that were probably vertical or nearly vertical as demonstrated by the present physical relationships at the block corners in the Clarks Fork Canyon and near Livingston (Figs. 5, 8). The north and east frontal edges of the mountain block rose to the highest structural position with relation to the adjacent, depressed block of the Bighorn Basin during Wasatch time. During Fort Union time, the covering sediments on top of the block were eroded (Stow, 1946). Accelerated erosion contributed large masses of debris to the Wasatch sedimentary rocks in the adjacent basin. In the vicinity of the Wyoming-Montana line, the Wasatch sedimentary rocks contain abundant granite pebbles derived from the frontal fault scarp or from the already denuded range crest.

Uplift continued after deposition of the Wasatch sediments as indicated by sharply upturned and locally faulted beds along the east side of the Beartooth front. Along the Nye-Bowler zone, continued downward movement on the south side during deposition of Fort Union beds created thicknesses up to 7600 feet near Red Lodge in contrast to half that thickness north of the fault zone (Wilson, 1936).

Left-lateral displacement of the zone during late Fort Union and early Wasatch time created asymmetric folding and en echelon faulting in the overlying rocks (Wilson, 1936). The tear faults, imbrication, lateral shearing and most of the structural detail along the Beartooth front developed during late Fort Union and Wasatch time.

During early uplift (Fort Union) Laramide porphyry sills and stocks intruded favorable structural zones, particularly along the Nye-Bowler zone, the Cooke City zone, and the east and northeast sides of the massif. These porphyry intrusions are absent in the mass of the Beartooth block between the Cooke City zone and the high frontal edge of the massif facing the plains. Most of the porphry intrusions predate frontal faulting as shown by their relationships with these faults along the northeast side of the block.

During the early Eocene, probably immediately following culmination of the Beartooth block's rise, and possibly triggered by it, volcanism and extrusion of the early acid and early basic breccias began in the general Absaroka-Yellowstone area. A series of vents roughly aligned along the Cooke City zone generally bound the northern side of the large volcanic-rock-covered region. The volcanic rocks effectively obscure the southwestern part of the Beartooth block. Considering the tremendous volume of extrusive rocks in the broad Absaroka-Yellowstone area, the southwestern part of the block as well as the area adjacent to it on the south may conceivably have been depressed by the massive transfer of subcrustal material to the surface. This could account for the nearly flat or continuing southwest dip of those Precambrian and early Paleozoic rocks exposed southwest of the Cooke City zone in the area southeast of Cooke City.

Relationships between the intrusive porphyries and the extrusive rocks along the northwestern part of the Cook City zone suggest that in places they are partly contemporaneous. However, intrusion of the porphyries began earlier (Paleocene) (Rouse *et al.*, 1937), and extrusion of volcanic rocks continued later Eocene) (Parsons, 1958).

Development of the great horizontally displaced masses of Paleozoic rocks described by Pierce (1957) at Heart Mountain, Sugarloaf Mountain, and elsewhere south of the Clarks Fork of the Yellowstone River began in the early Eocene, coincident with the ejection of large quantities of fragmental debris. The combination of extensive volcanism, probably with considerable seismic activity, and the force of gravity probably effected detachment of large areas of the Paleozoic rocks and movement of these masses downward along bedding planes and erosional surfaces. The mechanism of partial hydrostatic support Hubbert and Rubey (1959) proposed may have been involved in the movement of these masses.

The widespread normal faulting which characterizes Mio-Pliocene deformation in much of the Cordilleran area is noticeably lacking in the Beartooth area. A few small normal faults that are unrelated to prior structures have been mapped in the sedimentary rocks; these are relatively insignificant in terms of the block uplift. The bulk of later faulting seems to be associated with the Yellowstone Plateau and faults cutting the volcanic rocks.

By middle Eocene time the Beartooth block had attained its present full structural relief with respect to adjacent crustal blocks. However, it did not have its present topographic relief with respect to these areas. Faunal and floral evidence (Van Houten, 1956) from the Eocene rocks of the Bighorn Basin indicates that the entire area was at considerably lower elevations and enjoyed a mild, Savannah-type climate. During Miocene-Pliocene time, regional uplift of the Middle Rocky Mountains occurred in the order of 4000 to 6000 feet. Subsequent erosional modification created the present topographic relief.

Mackin (1937) has postulated some uplift of the Beartooth during Pleistocene time. His evidence is based on the elevation of stream terraces in the region. However, the same evidence can be interpreted in terms of glacial history without involving uplift.

TECTONIC INTERPRETATION

The detailed structural relationships of the perimeter of the Beartooth block have been described both spatially and chronologically. These details provide knowledge concerning local structural mechanisms. Now it is necessary to describe the over-all deformation of the block in terms of the interrelated structural mechanisms.

In the Beartooth Mountain block there are two distinct types of corner, one characterized by vertical uplifts, the second by vertical uplift with associated horizontal displacement. The vertical-lift corners of the Beartooth range are the northwest, or Livingston corner (Fig. 8), and the southern, or Clarks Fork corner (Fig.

5). At both corners a zone of frontal disturbance continues along the strike of the mountain front beyond the corner of the massif. Basement rocks exposed in the frontal continuation are separated from the more uplifted block by a near-vertical fault that strikes approximately perpendicular to the range front. In the northwest the Deep Creek fault in the Yellowstone River valley separates the tilted fronts of the Gallatin and Beartooth ranges, and in the south the Clarks Fork fault separates the Beartooth Mountains from the tilted ramp of Dead Indian Hill. Hence, the upward tilting of basement rocks beyond the corner of the massif provides a continuation of the basin edge and serves to confine the block corners against such lateral displacement parallel to the range front. Where tilting of one or several blocks is involved, cross faulting is necessary to free the end of the block for independent rotation. The significance of the vertically lifted block corners is that they firmly tie the mountain uplift to the adjacent basins and hence limit the amount of possible thrust displacement on the second type of corner, such as that at Red Lodge, where there is some associated horizontal displacement.

Outward from the vertical-lift corners and around the entire massif, great structural relief persists with respect to the adjacent basins. In the Beartooth Mountains this structural relief is commonly 2–4 miles.

The total structural relief may take place over a broad ramplike rise of the basement as in the case of the western Beartooth Mountains. Associated with the massif are a number of fault blocks tilted more steeply than the general angle of rise of the mountain fronts. Where measurable, the basement faults permitting this rise are nearly vertical or high-angle reverse faults. These faults have an opposite sense of displacement from that of the over-all rise of the mountain front. Some authors have called these structures underthrusts, but the present authors believe that the steep fault dips and the difficulty of simultaneous underthrusting and uplift of a mountain range make this interpretation unlikely. Instead, they interpret these as rotated blocks in which the rotations are the result of couples (in the vertical plane) produced by the differential uplift of the range with respect to the basins.

The parts of the mountain front along which the greatest structural relief occur are those characterized by frontal faults and thrusts. This is more than just a coincidence: where the greatest structural relief is produced along the mountain front, the mountain mass has its least confinement and is free to move outward for greater distances over the adjacent sedimentary rocks. Where observable, the frontal thrust dips 35° near Gardiner, 30°–40° near Red Lodge, and is almost flat at the Wyoming State line. Southward into Wyoming this fault can be seen to steepen and become vertical with increasing structural depth as it approaches the Clarks Fork corner. Commonly associated with the frontal thrusts are a series of imbricated thrust slices in the overturned veneer of sedimentary rocks. The sequence of sedimentary units within any single slice is almost invariably one of decreasing age outward from the mountain. The implication is that the thrust slices were developed in sedimentary rocks already steeply upturned against the mountain front.

At the Red Lodge corner of the Beartooth block there was initially a large change in strike of the bounding Beartooth fault zone. Uplift of the massif, lack of confinement, and outward movement along thrust and tear faults has resulted in maximum horizontal displacement of the massif at the Red Lodge corner.

Displacement of the mountain front by tear faults occurs along that part of the front with the greatest structural relief, the least confinement, and, hence, the greatest amount of horizontal displacement. The unconfined corner near Red Lodge is characterized by extensive tear faulting. The over-all displacement along the group of tear faults at the Red Lodge corner permitted certain parts of the mountain front to be moved outward farther than adjacent parts. Because there has been thrusting in two directions around the Red Lodge corner of the range and consequently greater resistance at the extreme corner, two major tear faults (Fig. 6; Pl. 3, fig. 2), each of them perpendicular to the range front, separate the corner block in keystone fashion and have permitted its less advance. It is significant that the maximum imbrication of frontal sedimentary rocks occurs along the front of this "keystone," or recessed block. The mechanism of imbrication as well as horizontal transport along the tear faults afforded release from compressive stress.

Offset along the frontal tear faults may be as great as 2 miles, implying a minimum horizontal displacement of 2 miles along the frontal faults. Part of this displacement may be more apparent than real because of differential uplift and erosion of the thrust and torn blocks. Additional field evidence is needed to define this possibility.

Figure 1. Aerial view (looking east). South-dipping Paleozoic rocks are in Dead Indian Hill (right), overlying Precambrian crystalline rocks exposed on both sides of the canyon. The Bighorn Basin in the distance. Photograph by Richards, Cody, Wyoming

Figure 2. Escarpment of the Clarks Fork fault. The Beartooth Plateau in the background is raised 3000 feet above the level of the sub-Cambrian crystalline rock surface at the Clarks Fork Canyon (foreground). View is north from Dead Indian Hill.

CLARKS FORK CANYON OF THE YELLOWSTONE RIVER, WYOMING

FOOSE, WISE, AND GARBARINI, PLATE 2
Geological Society of America Bulletin, volume 72

Figure 1. Line Creek fold, on the Montana-Wyoming line (looking west). Paleozoic sedimentary rocks upturned along the mountain front were folded along a vertical axis by right-lateral movement of Beartooth Mountain massif with respect to the Great Plains (foregroud).

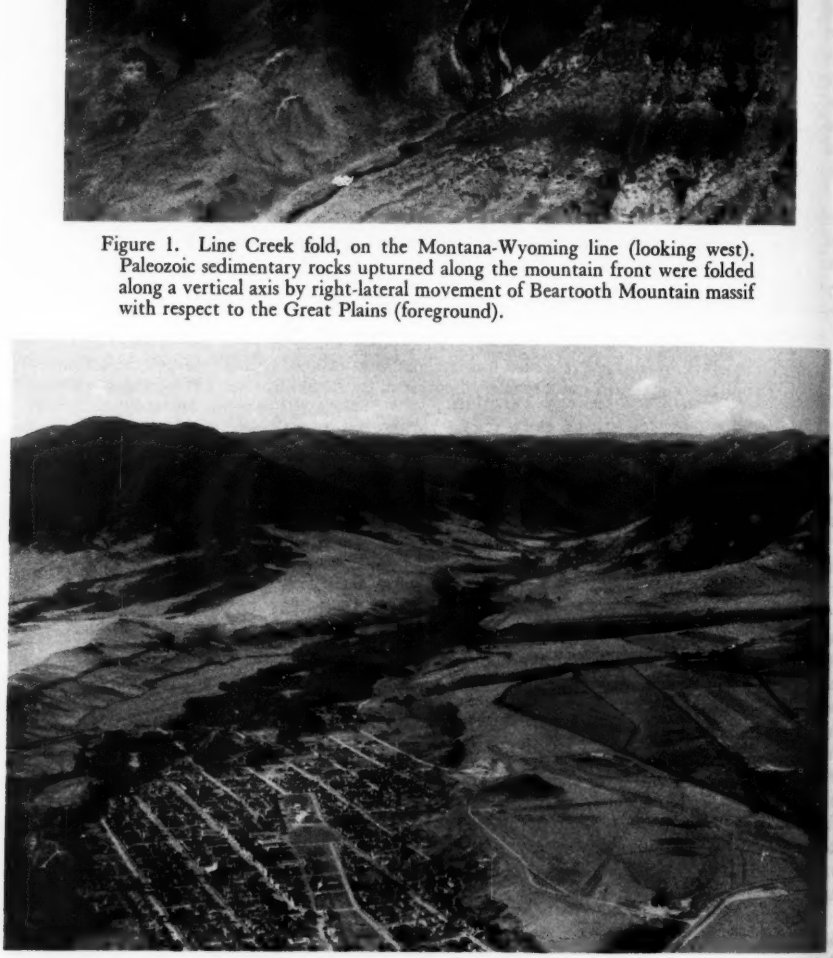

Figure 2. Aerial view of the Red Lodge corner of the Beartooth Mountains (looking southwest). Mt. Maurice is to the left of the Rock Creek valley which extends through Red Lodge (foreground). The mountain front (left) is displaced toward the Great Plains 10,000 feet along the Maurice tear fault with respect to the mountain front (right). Photograph by Richards, Cody, Wyoming

STRUCTURES ALONG THE BEARTOOTH MOUNTAIN FRONT, MONTANA AND WYOMING

FOOSE, WISE, AND GARBARINI, PLATE 3
Geological Society of America Bulletin, volume 72

Associated with tear faulting and horizontal movement toward the unconfined corner of the range at Red Lodge are the frontal lateral shear structures at Line Creek and near Red Lodge Creek (Fig. 7; Pl. 3, fig. 1). The structures are about 9 miles south and northwest respectively from the unconfined Red Lodge corner. Both structures are modifications superimposed on sedimentary rocks already upturned at the mountain front and, therefore, represent movement of the block under continuing compressive force during the late stages of uplift. The movement in both areas involved longitudinal movement of the crystalline mass of the block toward the unconfined corner, which created a massive lateral shear with associated faulting in the adjacent sedimentary rocks.

In summary, the Beartooth block has developed by dominantly vertical movement along frontal structures following early horizontal compressive and gentle folding of the region. Where vertical displacement became abrupt and great enough so that there was lack of buttressing against horizontal compression, horizontal displacement with associated imbrication and tear faulting resulted; the greatest thrusting occurred at the unconfined corner of the range at Red Lodge. Thus horizontal movement is only a later stage of the vertical uplift of the range, and a modification of the basic pattern of vertical rise as shown at the confined range corners.

In the Beartooth region tectonic speculations and theories have been almost as numerous as investigators.

As a method of block uplift, the Wedge theory of Chamberlin (1925) is commonly cited (Thom, 1955). With this mechanism, horizontal crustal compression on a downward-pointing wedge would cause the wedge block to rise along thrust faults bounding both edges. Thom (Personal communications, 1950; 1954; 1955; 1956; 1957; 1958) also favors the theory of rise of the range by movement along a ramp. Thom considered the ramplike thrust plane to flatten at depth, sled-runner fashion. This concept necessitates horizontal displacement of the mass to achieve range uplifts.

Associated with any statement of horizontal displacement in the Beartooth Region is the Heart Mountain problem (Pierce, 1957). Great horizontal displacement is indicated by blocks of Madison limestone resting on Tertiary sedimentary rocks of the Bighorn Basin many miles from any mountain front. Considerable confusion has resulted in the past from failure to distinguish between this feature of surficial sedimentary-rock displacement and that of fault displacement of the basement blocks. Pierce (1957) shows that the movement of these blocks was closely associated with some of the early volcanic activity in the Yellowstone-Absaroka region and thus represents a detachment thrust, aided at least in its later stages by gravity. The present authors likewise regard the Heart Mountain displacement as a later event associated with these early volcanic rocks and the shifting of a part of the sedimentary skin of the Beartooth core. Hence, the Heart Mountain structures should not be used as evidence for the amount of horizontal displacement that has occurred along the frontal faults of the major uplift.

The geologist willing to speculate on the deeper forces and mechanics of orogeny finds himself in an almost virgin field, barely cluttered by geologic facts! A full discussion of Laramide orogeny cannot be made here, but the detailed field relationships discussed provide a basis for certain ideas concerning the nature of deformation of the Beartooth Mountains that seem to be the most compatible with their orogenic history and the observed geologic facts. Thom (1923) stated some of these same basic ideas. The tectonic ideas of the present authors may be summarized as follows:

(1) Precambrian structural lines of weakness seem to have outlined both large and smaller units of the basement which reacted to the crustal forces of the Laramide deformation. This seems to have been true for a major portion of the Middle Rocky Mountain region.

(2) At the beginning of Laramide time, the dominant force in the crust probably was from the west or southwest. Even though many exceptions may be cited, evidence indicates that in its early stages, deformation in the crust of western North America "advanced" generally toward the east.

(3) With the application of compressive force on the crust of the Yellowstone-Bighorn region, the Beartooth block began to rise, and adjacent areas, like the Bighorn basin and the Crazy Mountain basin, were depressed. The writers suggest that initially this area was gently folded and that uplift and depression of crustal blocks, probably influenced by elements of basement anisotropism, followed.

(4) The writers suggest that plastic flow, involving the currentlike transfer of subcrustal material, became operative after the incipient rising, sinking, or tilting of blocks. Movement

would, of course, have been from beneath sinking blocks toward and under adjacent rising blocks. Hydrostatic pressure supported the blocks and also, possibly, caused the blocks to rise higher. Field relationships suggest that crustal movement was dominantly vertical; a large horizontal component of movement must have been operative simultaneously in the deeper region of plastic flow.

(5) When the rising Beartooth block, bounded mainly by vertical or very steep faults, attained considerable structural relief, none of the sides of the upper part of the block were supported or confined. Apparent spreading of the block began as a result of lack of confinement. Thereafter a dominant force along the frontal edge of the block was gravity. Continuing compressive forces from the southwest through the Beartooth block contributed to the major thrusting along the east and northeast sides of the block. Logically, the upper part of the vertical bounding faults were modified by newly developing thrusts with varying angles of dip toward the massif.

(6) Erosional shifting of material from the uplifted block increased the isostatic rise of the block, acting in addition to the main forces of compression and the pressure created by plastic flow.

(7) Throughout the developing deformation of the block, the overlying veneer of sedimentary rocks behaved in response to the movements of the basement below, arching, draping, breaking, slumping, and sliding, especially along and near the high, frontal edge of the block. These rocks contributed very little to the deformation of the block itself but were disrupted by their response to the movement of the block. Inevitably gravity played an important role in controlling the final resting place of erosionally detached masses and structurally broken pieces of the sedimentary veneer.

(8) The writers suggest that the Laramide orogenic movements terminated when east-west compression ceased and/or when the Beartooth block could no longer be raised hydrostatically.

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MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JUNE 29, 1959

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Cenozoic Stratigraphy and Structural Geology, Northeast Yellowstone National Park, Wyoming and Montana

Abstract: Cenozoic eruptive rocks and terrestrial sediments were deposited on a structurally complex basement of Precambrian, Paleozoic, and Mesozoic rocks in the Yellowstone Park region. Eruption of lower Cenozoic andesitic breccias and tuffs appears to have been cyclic, for the pyroclastic rocks become more mafic up-section with an increase in the relative abundance of breccia beds and decrease in tuff units. As each cycle proceeded, less silicic and more mafic material was ejected. The writer suggests that the cycles represent periodic tapping and drawing off of an upper, volatile-rich fraction of the source magma. A major hiatus separates the lower Cenozoic rocks from upper Cenozoic rhyolites and associated rocks in northern Yellowstone Park which are believed to be Pliocene and younger.

At least five phases of deformation can be recognized in the map area: (1) Precambrian metamorphism, (2) Lancian to early Eocene Laramide folding and faulting, (3) middle Eocene fracturing and normal faulting, (4) post-middle Eocene prelate(?) Pliocene normal faulting, and (5) post-late(?) Pliocene normal faulting of rhyolitic welded tuffs.

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INTRODUCTION

Geographic Setting

Yellowstone National Park is chiefly a forested volcanic plateau, 3348 square miles in area and 7000–9000 feet above sea level (Pl. 1). Mountain ranges rim three sides, rising 2000–5000 feet above the plateau. On the west, the plateau drops abruptly 1000 feet to the Snake River plain.

This report concerns 430 square miles of maturely dissected terrain in northeast Yellowstone Park. The region is drained by the Yellowstone and Lamar rivers and their many tributaries. Northeast Yellowstone Park and the adjacent country to the east offer unique contrasts in topography where the alpine sculptured Absaroka Range adjoins the youthfully dissected planation surface of the Beartooth uplift (Pls. 2, 6).

Between the Beartooth-Absaroka terrane and the northern edge of the Yellowstone rhyolite plateau, the country is maturely eroded, glaciated, and characterized by rolling hills and broad valleys. Locally the rivers course through steep-sided canyons. Soda Butte Creek valley and its tributaries are typical of the ice-sculptured high Absorokas. Postglacial landslip debris derived from high ridges of Cenozoic breccia breaks the U-shaped contour of some of these valleys and locally dams drainage courses.

Previous Work

Hague *et al.* completed reconnaissance geologic mapping of the Yellowstone region in 1896. Jones and Field (1929) and Howard (1937) conducted geomorphic studies in the Yellowstone valley. Howard focused considerable attention on stratigraphic relations of upper Cenozoic rocks in the Tower Junction area. Wilson (1934a; 1934b) investigated structural and stratigraphic problems in the Gardiner, Montana, area. B. E. Macomber (1956, M.A. thesis, Northwestern Univ.) described the geology of the Soda Butte area. Hamilton (1959) speculated on the petrogenesis of the Yellowstone rhyolitic rocks, and F. R. Boyd (1961) studied the petrology of the plateau rhyolites in detail. Paleobotanical studies are being conducted on the fossil floras of the Absaroka volcanic rocks by Erling Dorf (1960).

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Special thanks are due Yellowstone National Park personnel for hospitality accorded the writer and his colleagues in the field. The writer was assisted in the field by Messrs. E. Hanson, R. Burtner, and R. Lowe (1955) and J. Whetten, R. Manning, T. Sternberg, R. Dorf, and J. Cook (1956). Edward Shover assisted Professor Dorf in identifying fragmentary fossil plants from the Crandall(?) conglomerate, and B. E. Macomber kindly granted permission to include his findings from the Soda Butte area in this report. Mr. A. S. Pearce, Socony Mobil Oil Co. Inc., Dallas, Texas, drew the topographic sketch map (Pl. 6).

STRATIGRAPHY

General Statement

Precambrian, Paleozoic, Mesozoic, and Cenozoic rocks total more than 13,800 feet in the map area. Precambrian gneisses and schists of the Beartooth Mountains crop out as a high plateau, the margin of which is deeply dissected by south-flowing tributaries of the Yellowstone and Lamar rivers. Lower Paleozoic marine rocks and Cenozoic volcanic ejecta lie unconformably on the plateau surface and lower terraces. The Precambrian rocks of Buffalo Plateau and the highlands west of Hellroaring Mountain are intensely contorted and fractured. Near Gardiner, Montana, they are thrust over upper Cretaceous sandstones and shales. In the eastern part of the map area volcanic ejecta of the Absaroka Range cover the Precambrian rocks.

The Paleozoic sedimentary rocks are chiefly

marine carbonates and shales. They include the middle and late Cambrian Flathead, Wolsey, Meagher, Park, Maurice, Snowy Range, and Grove Creek formations. These, in turn, are overlain by the Ordovician Bighorn formation, the Devonian Jefferson formation, and the Mississippian Madison group limestones. Pennsylvanian Amsden(?) beds crop out in an isolated fault block 7 miles southeast of Gardiner, Montana, at Tiger Hill.

Clastic marine and terrestrial rocks characterize the Mesozoic section in the Mount Everts area. On the west face of Mount Everts, coarse clastics of the Mesaverde group overlie the Mowry, Frontier, and Cody formations. Isolated outcrops of the Ellis and Morrison(?) formations were found in the fault complex with Amsden(?) strata.

Cenozoic rocks in northern Yellowstone Park consist mainly of ash, tuff, and breccia associated locally with basaltic lavas and fluvial gravels. Members of early field parties (Hayden, 1872; Jones, 1875) recognized the volcanic rocks of the Yellowstone-Absaroka region. Hague *et al.* (1896) subdivided the volcanic rocks in terms of distinctive petrographic characteristics, and his names (early acid breccia, early basic breccia, *etc.*) have become entrenched in the literature (Fig. 1). Although a formal system of stratigraphic nomenclature should be established for these rocks, this should evolve from regional studies encompassing all the Yellowstone-Absaroka region rather than from a local study such as this one. Use of Hague's nomenclature, plus terms introduced by Howard (1937) and Boyd (1961), does not restrict geologic interpretation of the rocks in question.

Measured sections (Pl. 1) have been filed with the ADI Auxiliary Publications Project, Photoduplication Service, Library of Congress, Washington 25, D. C.[1]

Nomenclature of the Pyroclastic Rocks

The writer follows Williams *et al.* (1954) in classifying pyroclastic rocks. R. L. Hay's (1952, Ph.D. thesis, Princeton Univ., p. xi–xii) criteria for differentiating silicic, mafic, and intermediate rock types were applied in the map area with slight modification. Terms describing compositional variations of the pyroclastic rocks follow.

[1]To obtain photoprints or 35-mm microfilm, cite doc. #6775, remit $1.25. Make checks payable to Chief, Photoduplication Service, Library of Congress.

Rhyolitic tuff: Chiefly glass shards with phenocrysts of quartz, sanidine, Na plagioclase, rare augite, magnetite, and zircon; 69–78 per cent SiO_2

Trachyrhyolitic tuff: Similar to rhyolitic tuff but free quartz is absent, sanidine is abundant; 64–69 per cent SiO_2

Quartz dacite porphyry tuff: Dacite crystal tuff with abundant quartz phenocrysts, common biotite, rare hornblende; +67 per cent SiO_2

Dacite tuff or breccia: Biotite-rich tuff or breccia with common or abundant hornblende, very rare pyroxene and quartz phenocrysts; 67 per cent SiO_2

Acid andesite tuff or breccia: Hornblende and biotite the most common mafic minerals, pyroxene rare but usually present; groundmass may contain rare quartz; 62–67 per cent SiO_2

Medium andesite tuff or breccia: Pyroxene the most abundant mafic mineral, some hornblende, very rare olivine; 56–62 per cent SiO_2

Basaltic andesite tuff or breccia: Pyroxene abundant, olivine common; plagioclase approximately An_{56}; abundant magnetite crystals occur in a dark brown groundmass; 52–56 per cent SiO_2

Sedimentary rocks with pyroclastic material incorporated from direct ashfall or reworking are called volcanic conglomerates, sandstones, and mudstones. The use of mafic minerals in correlation of pyroclastic rocks has limitations, owing to the susceptibility of these minerals to uneven subaerial distribution and postdepositional alteration. However, the gross composition of a bed could be determined by examination of 10–20 specimens from that bed, which permitted detection of uneven distribution of phenocrysts and differential weathering. More commonly, 5–10 specimen analyses sufficed.

Lower Cenozoic Rocks

Crandall(?) conglomerate. Pierce (1957) named the Crandall conglomerate from exposures on the south side of Crandall Creek, a tributary of the Clarks Fork of the Yellowstone, 12 miles east of Yellowstone Park. In the type area, the Crandall conglomerate is a coarse stream-channel deposit composed chiefly of pebbles, cobbles, and boulders of pre-Pennsylvanian carbonates and basement rocks. The type Crandall conglomerate is overlain by early acid breccia, and its structural and stratigraphic relations suggest that the deposit is early Eocene (Pierce, 1957, p. 613).

A massive gneiss-schist boulder conglomerate crops out in the Yellowstone River canyon between Hellroaring and Elk creeks (Pl. 1, measured section 1). The conglomerate beds are virtually flat-lying and are exposed in near-vertical walls, 109 feet high, for 1.3 miles upstream from the mouth of Hellroaring Creek.

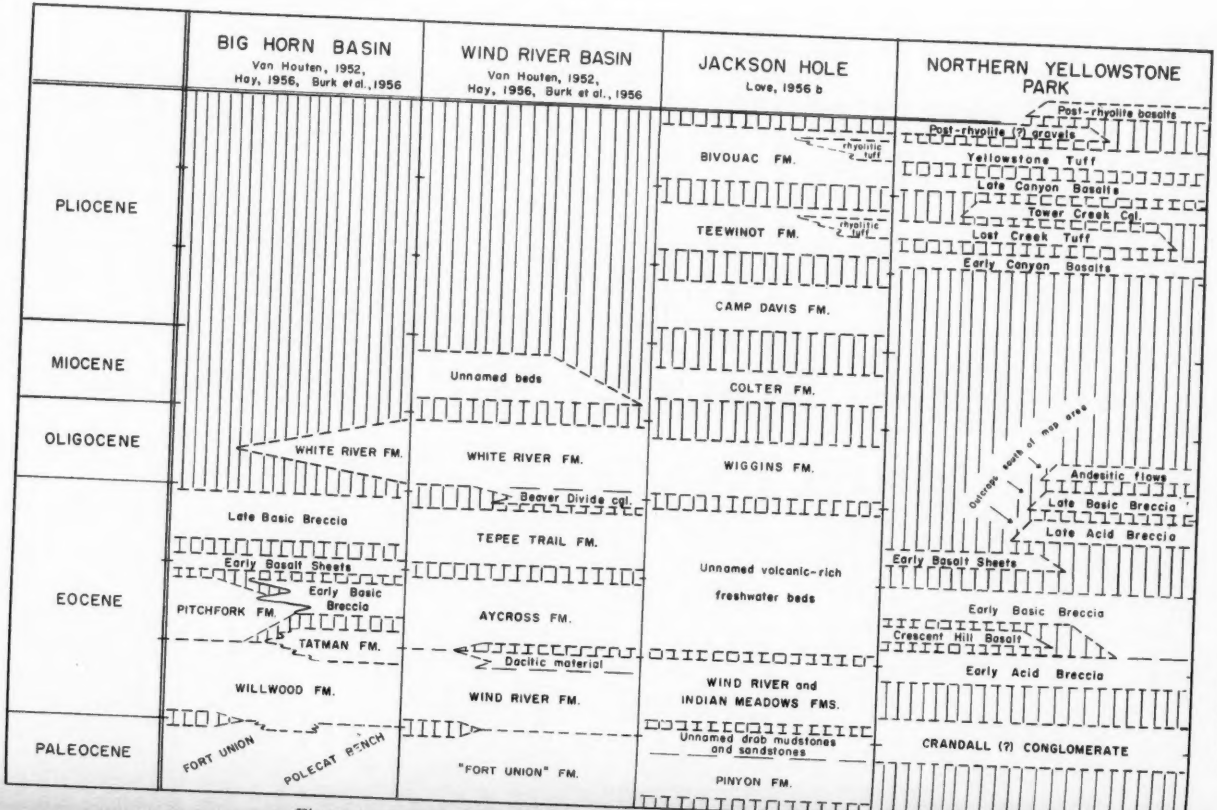

Figure 1. Cenozoic correlation chart, Yellowstone National Park region

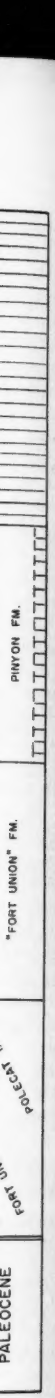

Figure 1. Cenozoic correlation chart, Yellowstone National Park region

These rocks appear to fill a channel cut in the Precambrian metamorphic rocks of the Beartooth uplift, and they are overlain unconformably by early acid breccia. The conglomerate consists of tan to gray, well-rounded gneiss, schist, and rare quartzite boulders in a tan to gray matrix of coarse arkosic wacke. Some boulders are as large as 8 feet in diameter. Sparse thin sandstone lenses and silty micaceous shales were seen in the deposit. Volcanic materials were conspicuously absent. The formation was not observed elsewhere in the map area.

The writer correlates this conglomerate with Pierce's Crandall conglomerate to the east, suggesting genetic similarity and probable contemporaneity but not physical continuity. The Crandall(?) conglomerate in the map area is fractured: joint planes cut boulders and matrix alike. The overlying early acid breccia shows weak jointing parallel to the fractures in the Crandall(?) conglomerate and Precambrian rocks. Major joints in the conglomerate do not pass into the overlying volcanic rocks.

Fragmentary fossil leaves from the Crandall(?) conglomerate include *Chrysobalanus(?) lanceolatus* Knowlton (Paleocene), *Rhammus belmontensis* Knowlton and Cockerall (late Cretaceous to Eocene), and *Juglans rugosa* Lesquereux (Paleocene to Eocene). A large collection of fossil leaves from early acid breccia immediately above the Crandall(?) conglomerate is late early Eocene (Dorf, 1960).

Other coarse lower Cenozoic conglomerates in the Yellowstone Park region, found close to the margins of uplifted Laramide ranges, consist largely of material derived from the range cores and show some deformation (Peale, 1896; Love, 1947; 1956b; J. A. Mann, 1950, Ph.D. thesis, Princeton Univ.; Scholten *et al.*, 1955). Although the rocks near the mouth of Hellroaring Creek resemble these conglomerates, they are correlated specifically with Pierce's (1957) Crandall conglomerate for the following reasons. Both crop out on the flank of the Beartooth uplift and are overlain by early acid breccia. The conglomerate in the map area is almost certainly a stream deposit, as is the Crandall conglomerate. They are compositionally similar, and both units appear to be latest Paleocene or early Eocene in age (Pierce, personal communication, 1959).

Early acid breccia. Named by Hague *et al.* (1896), the early acid breccia is the oldest stratigraphic unit in the Cenozoic pyroclastic sequence of Yellowstone Park. These rocks are light-colored tuffs and breccias characterized by an abundance of biotite, quartz, and hornblende phenocrysts. They are finer textured and have a higher silica content than the overlying darker, more mafic, pyroxene-rich andesite breccias and tuffs which Hague called the early basic breccia.

The early acid breccia is typically exposed in the vicinity of Crescent Hill (Pl. 1, measured section 1) in a section about 1550 feet thick. The basal dacite breccia has common gneiss and schist fragments from the Precambrian basement complex. Fragments of Paleozoic rocks are less common, although they are abundant in the early acid breccia of the Crandall quadrangle east of Yellowstone Park (Rouse, 1937, p. 1264). The basal dacite breccia is overlain by a thick sequence of dacite crystal and vitric tuffs, lapilli tuffs, and sparse breccia beds. Quartz dacite porphyry tuffs are found locally. Acid andesite and dacite breccias are more common in the upper part of the Crescent Hill section than in the lower part. Porcellanites are common in the lower 200 feet and upper 400 feet of the section. Red-brown to black andesitic Crescent Hill basalt caps the Crescent Hill section. Other outcrops of early acid breccia are found at Specimen Ridge, in the Lamar River valley near the mouth of Slough Creek, and in Cache Creek valley (Pl. 1). A massive volcanic-boulder conglomerate, 10 feet thick, underlies quartz dacite porphyry crystal tuffs on the north side of the Lamar River, 0.25 miles above the mouth of Slough Creek. The conglomerate is composed of silicic volcanic rocks with lesser amounts of gneiss, schist, and quartzite. Thin cross-bedded plant-bearing sandstone beds occur in the middle of the section. The volcanic conglomerate is assigned to the early acid breccia rather than the Crandall(?) conglomerate because of the abundant volcanic material present.

One aim of the present investigation was to clarify the contact relationship between the early acid and early basic breccias. Hague *et al.* (1896, p. 6; 1899, p. 270) indicated that in some areas the contact was gradational and in others it was erosional. Rouse (1937, p. 1265) attributed the patchy distribution and variable thickness of the early acid breccia to widespread erosion before the early basic breccias were erupted. Hay (1956, p. 1885), from a brief reconnaissance of the Tower Junction area in Yellowstone Park, suggested that stratigraphic relationships of the early acid and early basic breccias appeared more complex than Hague

et al. (1896) indicated. He reported (p. 1885) that rocks of dacitic composition, mapped as early acid breccia, contained volcanic conglomerates of andesitic composition similar to those in the early basic breccia. Hay concluded that possibly all or part of the early acid breccia interfingers with early basic breccia and that the two units are equivalent to interfingering dacitic and andesitic beds in the Pitchfork formation of the southeastern Absaroka Mountains.

The writer mapped the Tower Junction area in detail and found volcanic conglomerate (dacitic composition) at only one locality in the early acid breccia. This is the conglomerate noted earlier along the Lamar River near the mouth of Slough Creek. Andesitic pyroclastic material was found in the dominantly dacitic sequence of the early acid breccia. However, these rocks are the acid andesite eruptive facies common to the upper part of the early acid breccia throughout the map area. Hague *et al.* (1899, p. 270, 271, 274) recognized these rocks. Westward from the summit of Crescent Hill along the old Mammoth-Tower road the early acid breccia ranges from quartz dacite porphyry to dacite to acid andesite in composition upsection. The acid andesite breccias and tuffs (intermediate pyroclastic facies) are lithologically similar to a "transitional lithology" Rouse (1937, p. 1265) described in the upper part of the early acid breccia in the Absaroka Mountains east of Yellowstone Park.

The writer observed the contact between the early acid breccia and overlying early basic breccia at six localities (Pl. 1).

At the base of a ridge separating Blacktail Deer Creek drainage from Geode Creek drainage, 2.6 miles N. 10° W. from Prospect Peak, acid andesite breccias are conformably overlain by medium andesite and basaltic andesite breccias. The contact is sharp, occurring at the boundary between two distinct beds.

Two elongate outcrops of basaltic andesite to medium andesite breccia cap a divide of dacite to acid andesite tuffs and breccias between Elk Creek and Geode Creek drainage basins, 2.5 miles N. 20° E. from Prospect Peak (Pl. 3, fig. 1). This contact can be traced southeast for 1.4 miles, thence north and northeast down the unnamed creek valley between Elk Creek and the stream that flows to the head of Lost Lake meadow. The contact is apparently erosional in this area but is lost in the forested slopes of the lower valley. Hills on the sides of the valley are composed of flat-lying early acid breccia (northwest) and early basic breccia (southeast). The creek appears to flow along the contact between the two formations.

The contact between the early acid breccia and the early basic breccia is exposed at the northwest end of Specimen Ridge (Pl. 1, measured section 2). Interbedded quartz dacite porphyry, dacite, and acid andesite tuffs and breccias grade upward to predominantly dacitic beds which are conformably overlain by medium andesite breccias. A bedding-plane contact separates the two formations.

Approximately 500 yards southeast of measured section 2, on the northeast face of Specimen Ridge, the early acid breccia is in erosional contact with early basic breccia. Horizontal beds of medium andesite breccia abut against horizontal beds of acid andesite crystal tuff underlain by dacite vitric to crystal tuff.

Horizontal beds of medium andesite crystal tuff overlie quartz dacite porphyry crystal tuffs with erosional unconformity near Quartz Creek, 2.1 miles S. 55° E. from Tower Fall.

In Cache Creek valley, the contact is gradational from dacite and acid andesite breccia to medium andesite breccia in a vertical interval of 30 feet. Locally the contact is unconformable (B. E. Macomber, 1956, M.A. thesis, Northwestern Univ., p. 41–42).

These observations confirm the conclusions of Hague *et al.* (1896; 1899). The contact between the early acid and early basic breccias is conformable in some areas and unconformable in others. The writer could not find any field evidence to support Hay's (1956) suggestion of interfingering.

Fossil floras from the early acid breccia indicate a late early Eocene age for these eruptive rocks (Dorf, 1960). This suggests that the early acid breccia correlates with part of the Tatman formation, uppermost Willwood formation, and lower Pitchfork formation of the Big Horn Basin. In the Wind River Basin, dacitic pyroclastic material of late early Eocene age in the Wind River formation and the lower part of the Aycross formation appear to be equivalent to the early acid breccia of Yellowstone Park. Dacitic and acid andesite tuffs interbedded with claystones, sandstones, and conglomerates 3 miles south of Togwotee Pass represent the first Eocene volcanic activity in the Jackson Hole area and may be contemporaneous with the early acid breccia to the north (Love, 1956b, p. 85).

Crescent Hill basalt. Howard (1937, p. 18) renamed the basalts overlying early acid breccia

in the Crescent Hill area the Crescent Hill basalt. Hague *et al.* (1896) originally mapped these rocks as "post-rhyolite Recent basalt." The basalt is typically exposed at the summit of Crescent Hill.

The Crescent Hill basalt, a fine-grained to porphyritic andesitic basalt flow as much as 60 feet thick, crops out as erosional remnants capping hills of early acid breccia. The flow differs from other basaltic lavas in the area by being scoriaceous to pumiceous in the upper part and commonly sheared and silicified.

Howard (1937) regarded the Crescent Hill basalt as older than the rhyolites because it shows more extensive deformation. The writer concurs and believes that the position of the Crescent Hill basalt in the stratigraphic column can be further defined using the amount of deformation as a criterion. Study of aerial photographs of the Crescent Hill area indicates a prominent rectangular joint pattern in the early acid breccia and Crescent Hill basalt. Local fault and shear patterns in the basalt parallel this joint system. The joint system is not seen in adjacent outcrops of early basic breccia and the Yellowstone tuff. Consequently, the writer regards the Crescent Hill basalt as definitely post-early acid breccia and probably pre-early basic breccia.

Early basic breccia. Red-brown and dark-gray medium andesite and basaltic andesite breccias and tuffs constitute the major part of the Absaroka and Washburn ranges in northern Yellowstone Park. Locally, these deposits contain volcanic conglomerates and sandstones. Hague *et al.* (1896) named these rocks the early basic breccia.

In the map area, the early basic breccia is chiefly layered medium andesite breccia and tuff which becomes more mafic up-section. Unusually fine exposures can be seen at Baronett Peak in Soda Butte Creek valley (Pl. 3, fig. 2). Coarse breccias are more common in the early basic breccia than in the underlying early acid breccia. Basaltic andesite lavas are commonly intercalated with basaltic andesite breccias and tuffs in the upper part of the early basic breccia. Contorted basal beds in Soda Butte Creek valley suggest possible mudflow origin for that part of the section, but they may only indicate local slumping.

Regionally, the early basic breccia attains a maximum thickness of 5500 feet near Dead Indian and Trout peaks east of the map area (Rouse, 1937). Locally, the early basic breccia is 3000 feet thick at The Thunderer on the southeast side of Soda Butte Creek valley (Pl. 1). At Fossil Forest in the Lamar River valley (Pl. 1, measured section 4), the early basic breccia consists of about 1400 feet of medium andesite breccias and tuffs overlain by 250 feet of basaltic andesite breccia and tuff. The basaltic andesite facies continues another 1000 feet above the Fossil Forest section to the summit of Amethyst Mountain: the coarse pyroclastic rocks there resemble the chaotic debris which forms the bulk of Mount Washburn 8 miles to the southwest. The early basic breccia is overlain locally by the early basalt sheets and younger rocks.

One of the more remarkable chapters of earth history is exposed for the trained reader at Fossil Forest. Periodic breccia and ash eruptions buried successive forests during Eocene time; the forest remains are now silicified and exposed on the steep slopes overlooking the Lamar valley. Jones (1875) made brief mention of abundant fossil wood near Specimen Ridge, and Holmes (1883) brought the fossil forests of the Lamar valley to scientific attention. Knowlton (1921) described the section in detail, reporting 12 forest levels. The size of the entombed pines and sequoias suggested to Knowlton that each forest level required 200 to 1000 years to attain maturity. He estimated (p. 27–28) that the forests flourished for 2400 to 12,000 years.

The writer measured the Fossil Forest section in detail and recorded 27 forest levels (measured section 4). Criteria for determining a given forest level in the breccia and tuff sequence included the expansion of the base of an upright tree trunk immediately above the root system and an abundance of fallen logs at a given stratigraphic level. This writer suggests that the Fossil Forest section represents 5400 to 27,000 years of intermittent volcanic eruption and forest development.

Fossil leaves from the early basic breccia indicate middle Eocene age (Dorf, 1960). This would make the early basic breccia equivalent to part of the Pitchfork formation in the Big Horn Basin and correlative with the Aycross and lowermost Tepee Trail formations of the Wind River Basin.

Early basalt sheets. A sequence of orthoclase-bearing porphyritic basalts 700–1000 feet thick overlies early basic breccia and forms the east wall of Mirror plateau in the southeast part of the map area. These rocks are also typically exposed on The Thunderer. Hague *et al.* (1896) named these flows the early basalt sheets.

Similar basalts of the Absarokite-Shoshonite-Banakite clan cap high peaks and divides in the northern Absaroka Mountains. These rocks can be traced southward to Pyramid Peak, east of Yellowstone Lake, where late acid breccia overlies them (Hague *et al.*, 1896).

Late acid breccia and late basic breccia. The eruptive rocks which Hague *et al.* (1896) mapped as late acid breccia and late basic breccia were not observed in the map area. South and east of the map area, they are overlain by andesitic flows and represent a second cycle of eruptive activity similar to that indicated by the early acid and early basic breccias in northern Yellowstone Park. Dorf (1960, personal communication) indicated that preliminary examination of fossil floras from the late breccias suggests late Eocene or early Oligocene age. In the Mink Creek area of northern Teton County, Wyoming, 3–5 miles south of Yellowstone National Park, rocks mapped earlier as late basic breccia (Hague *et al.*, 1896), but now included in the Wiggens formation, contain abundant early Oligocene (Chadron) vertebrate fossils (Love, 1956b, p. 88).

Upper Cenozoic Rocks

Discussion. Deposits which postdate the early breccia sequence, and presumably the late breccias, include basaltic lavas, rhyolitic and trachyrhyolitic welded tuffs, fluvial and lacustrine sediments, glacial till, landslip and hot-springs deposits, and valley fill. Preglacial rocks are difficult to correlate, in part, because of scattered outcrops, absence of fossils, and lack of distinguishing characteristics of the intercalated basaltic lavas. Correlations are based on similar lithologies and thicknesses, position in similar sequences, relative structural disturbance, derived conglomeratic material, and topographic position. Stratigraphic interpretation depends in part on geomorphic interpretation of drainage history in the Yellowstone and Lamar river valleys. Jones and Field (1929, p. 270–271) believed that all the "Recent basalts" postdated the plateau rhyolites because the basalts are apparently re-

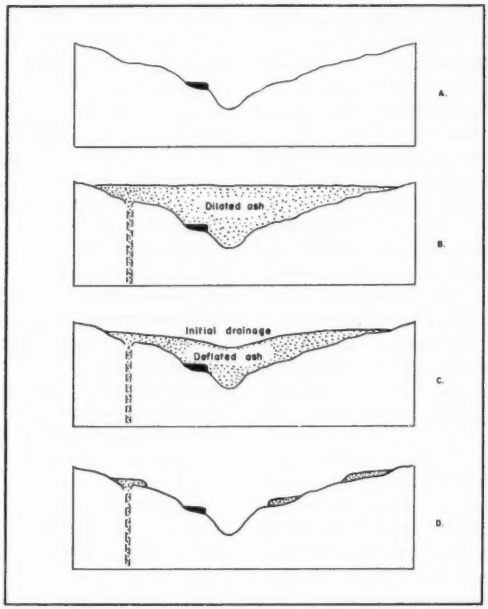

Figure 2. Exhumation of fossil valleys, Yellowstone National Park

Figure 3. Upper Cenozoic stratigraphic sections, Yellowstone River valley, Tower Junction area

stricted to the present valleys of the Yellowstone and Lamar rivers which are cut in the rhyolite plateau. Howard (1937) rejected the possibility of postrhyolite exhumation of the fossil Lamar valley. Boyd (1961, p. 396) advanced the hypothesis that present topography in parts of the Yellowstone river valley is exhumed pre-Yellowstone tuff topography. The writer arrived at the same conclusion in his area to the north of Boyd's area. Increased thickness of Yellowstone tuff along the present courses of Agate, Lost, Tower, Blacktail Deer, and Lava creeks in the map area suggests that drainage may have existed along these same courses prior to the eruption of the rhyolitic tuffs. As Boyd noted, compaction of the tuff deposit over subjacent topography (Fig. 2) would concentrate new drainage over the ancient valley axes. Physical correlations all point to a pre-Yellowstone tuff age for the basalts in the Tower Junction area, thereby supporting the "exhumation hypothesis."

Figure 3 indicates the upper Cenozoic sections examined in the Tower Junction area. Correlations relate the various outcrops to the excellent exposures in the Yellowstone River canyon near Tower Fall (Pl. 4).

Early canyon basalt. Hague *et al.* (1896) named the lower flows in the Tower Fall area the early canyon basalt. The flows are finely to coarsely crystalline, pyroxene- and olivine-porphyry basalts, about 30 feet thick in the canyon exposures. The lower part is columnar, and the upper part is characteristically blocky. In some places along the canyon the blocky part has been eroded away prior to deposition of the overlying Tower Creek conglomerate.

The fact that early canyon basalt is absent on the west side of the canyon at Tower Fall suggests that the flow seen on the east side lies on an old east-facing terrace cut in the early basic breccia. Terrace cutting probably occurred when the Yellowstone River occupied the now till-blocked canyon 0.5 miles east of its present course (Howard, 1937, p. 38), prior to aggradation of the valley that came with deposition of the Tower Creek conglomerate. At the northern end of the canyon exposures, early canyon basalt crops out at the same elevation on both sides of the river. On the east side the basalt is overlain by Tower Creek conglomerate which is absent or masked by glacial till on the west side. Here, the early canyon basalt is at the same elevation as a similar flow that crops out beneath Lost Creek tuff on the north side of Specimen Ridge, 1.5 miles southeast across the preglacial Yellowstone River valley. Howard (1937, p. 33) mapped the basalt near Specimen Ridge as Elk Creek basalt, but the writer believes that in terms of petrographic similarity, proximity, and topographic position in the ancestral Yellowstone valley the rocks are early canyon basalt. These relationships are important in that they fix the age of the Lost Creek tuff as post-early canyon basalt.

In the vicinity of Tower Junction, several other basalts are mapped as early canyon basalt largely in terms of their position beneath Tower Creek conglomerate or Lost Creek tuff. Petrographically, they all are similar to the typical early canyon basalt.

Lost Creek tuff. Scattered outcrops of buff- to yellow-weathering purple and gray welded tuff are found in the Tower Junction area (Pl. 1). Petrographically, the rock resembles the dacitic facies of the early acid breccia but is rich in sanidine and does not contain free quartz. Plagioclase phenocrysts are commonly decomposed, and biotite is very rare. The groundmass of the tuff consists chiefly of vitric shards with some pumiceous lithic fragments, and the shards have the characteristic eutaxitic structure of welded tuffs. A few pieces of charred wood were found in the tuff at one outcrop. The tuff is chemically intermediate between rhyolite and trachyte (Hague *et al.*, 1899, p. 324). It is typically exposed at Lost Creek meadow, at the south end of Garnet Hill, at the northeast base of Crescent Hill, and on the north side of Junction Butte. The tuff is 40–45 feet thick at maximum.

Originally, the Lost Creek tuff was called "trachytic rhyolite" (Hague *et al.*, 1896). Howard (1937) named the unit the Lost Creek trachyte, and the present writer has modified this to clarify the pyroclastic origin of the rock.

Hague *et al.* (1896) suggested that the "trachytic rhyolite" was younger than the early basic breccia and older than the early basalt sheets. Howard (1937, p. 25–26) indicated that the Lost Creek "trachyte" was younger than Elk Creek basalt and older than the plateau rhyolites. The writer agrees with this and would further define the age of the Lost Creek tuff, based on sections in the Tower Junction area, as post-early canyon basalt, pre-late canyon basalt, and possibly pre-Tower Creek conglomerate. Howard's Elk Creek basalt is regarded as part of the early canyon basalt sequence.

Tower Creek conglomerate. Hague *et al.*

Figure 1. Typical alpine topography of Absaroka Range, 6 mi. east of map area. Early basic breccia constitutes main mountain masses, lies unconformably on Paleozoic marine rocks. Matterhorn (center) is Pilot Peak; Index Peak immediately to right.

Figure 2. Youthfully dissected Precambrian surface of Beartooth Range between Cooke City and Red Lodge, Montana, 30 mi. east of map area. View northwest from main highway at an elevation near 11,000 feet

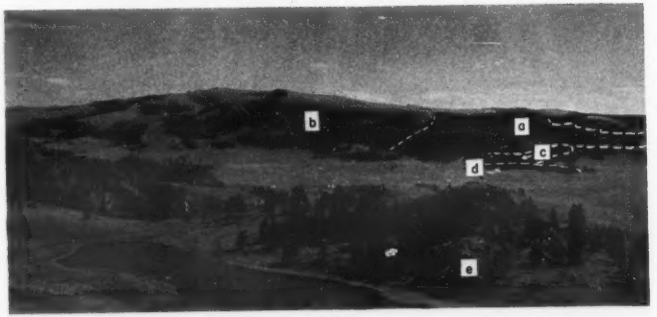

Figure 3. Specimen Ridge viewed from south flank of Buffalo Plateau (a) early acid breccia, (b) early basic breccia, (c) Lost Creek tuff, (d) early canyon basalt, (e) Precambrian gneiss

Figure 4. View southeast up Lamar River valley toward (a) Fossil Forest and (b) Absaroka Range

PHYSIOGRAPHY, NORTHEAST YELLOWSTONE PARK REGION

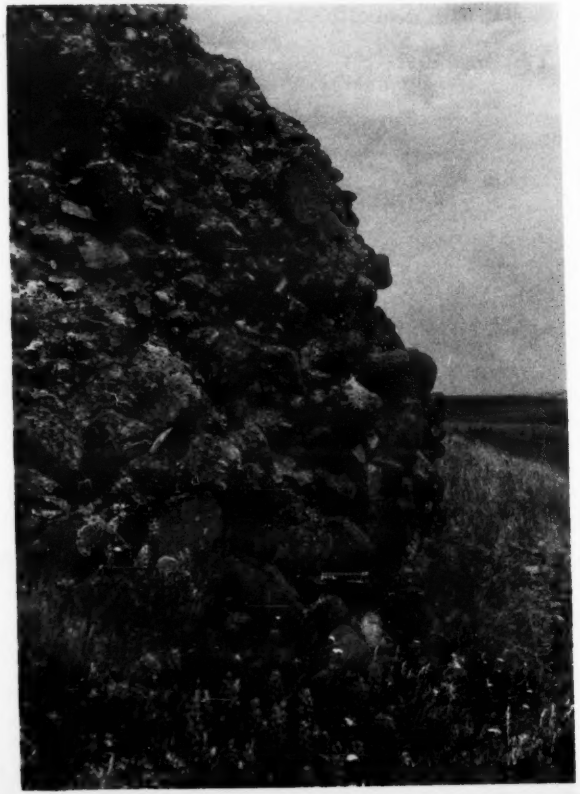

Figure 1. Basaltic andesite and medium andesite breccia, north flank of Prospect Peak

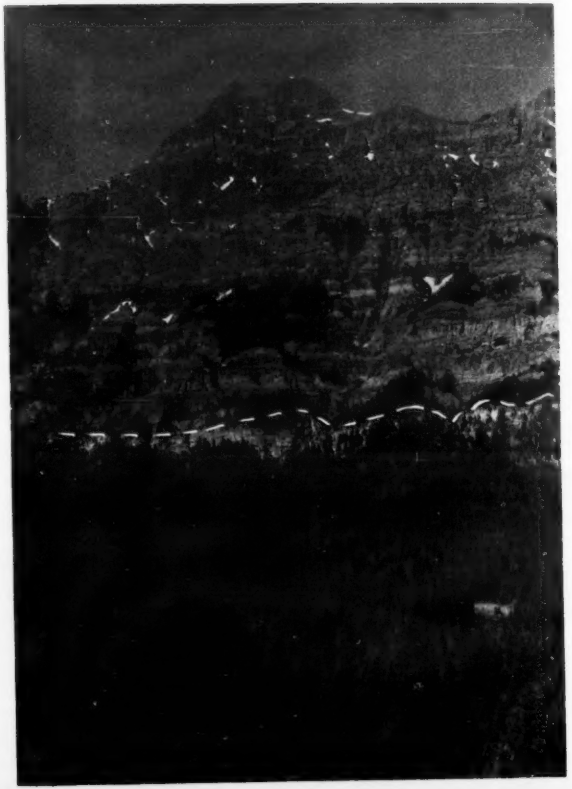

Figure 2. Baronett Peak viewed southwest across Soda Butte Creek valley. Layered early basic breccia unconformably overlies light-colored Mississippian carbonates of Madison group. Contact shown by dashed white line

Figure 1. Basaltic andesite and medium andesite breccia, north flank of Prospect Peak

Figure 2. Baronett Peak viewed southwest across Soda Butte Creek valley. Layered early basic breccia unconformably overlies light-colored Mississippian carbonates of Madison group. Contact shown by dashed white line

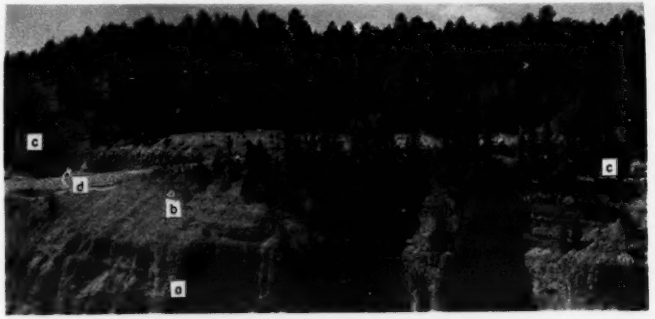

Figure 1. West wall of Yellowstone River canyon, 0.5 mile north of Tower Fall: (a) early basic breccia, (b) Tower Creek conglomerate, (c) late canyon basalt, (d) outcrop shown in Figure 4, this plate

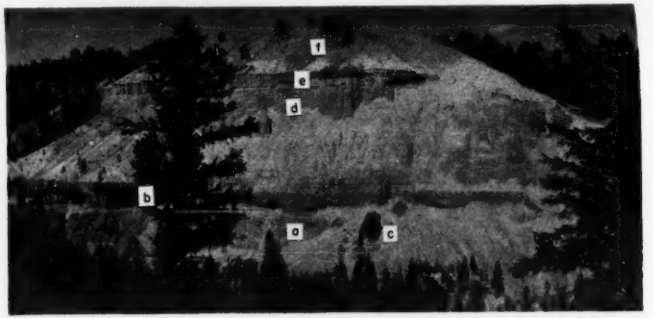

Figure 2. East wall of Yellowstone River canyon, viewed from Tower Fall: (a) early basic breccia, (b) early canyon basalt, (c) slump block of early canyon basalt, (d) Tower Creek conglomerate, (e) late canyon basalt, (f) glacial till

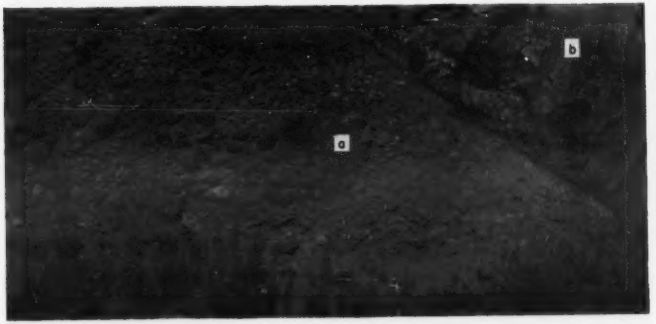

Figure 3. South end of Overhanging Cliff member of late canyon basalts, 0.25 mi. north of Tower Fall: (a) horizontal Tower Creek conglomerate, (b) Overhanging Cliff basalt filling channel cut in conglomerate

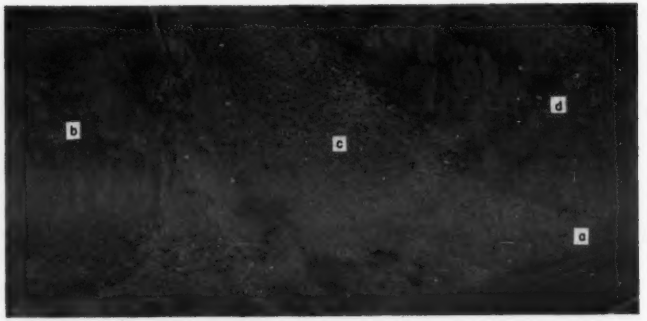

Figure 4. North end of Overhanging Cliff member of late canyon basalts, 0.5 mi. north of Tower Fall: (a) Tower Creek conglomerate, (b) Overhanging Cliff basalt, (c) fossil talus from Overhanging Cliff member, (d) younger flow of late canyon basalt sequence

STRATIGRAPHIC SECTIONS, YELLOWSTONE RIVER VALLEY, TOWER FALL AREA

Figure 1. Base of 150-foot cliff of columnar welded tuff. Lower 17 feet of section highly vesicular. Vesicles flattened to approximately 0.5 by 1.5 inches

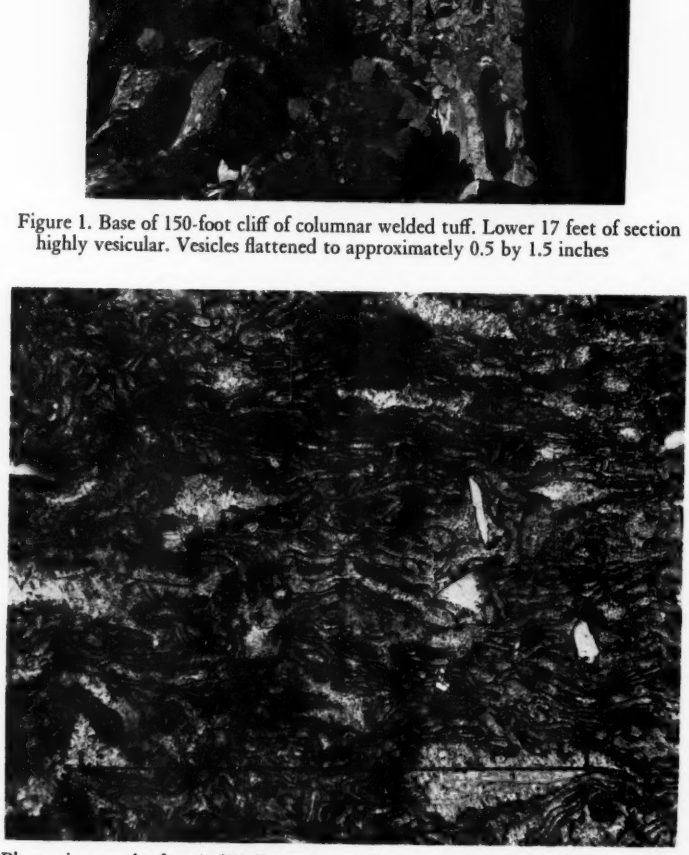

Figure 2. Photomicrograph of typical Yellowstone tuff showing annealed and flattened shards, pumice fragments, and quartz phenocrysts: plain light

YELLOWSTONE TUFF, LOST CREEK VALLEY

(1896) named fluvial gravels in the walls of the Yellowstone River canyon near Tower Fall "canyon conglomerate." Hague *et al.* changed the name to Tower Creek conglomerate in 1904 (Wilmarth, 1938, p. 2173).

The Tower Creek conglomerate is a friable boulder and cobble deposit consisting largely of fragments of Precambrian gneiss, schist, rare Paleozoic(?) quartzites, and some volcanic material. The conglomerate ranges in thickness from 150 feet on the east side of the canyon to 40 feet on the west side. Farther downstream, at Baronett Bridge, the unit is 25–30 feet thick. Howard (1937, p. 44, 51–53, 62) reported fragments of rhyolite in the Tower Creek conglomerate, but the writer and his colleagues failed to find rhyolitic pebbles in these beds on the east and west sides of the canyon. The pyroclastic rocks noted were similar to material in the early acid and early basic breccias. For the most part, the pyroclastic rocks were deeply weathered, and specific identification was impossible. Hague *et al.* (1896, p. 3) also reported that in spite of careful search they failed to find rhyolitic pebbles in the conglomerate. This discrepency between Howard's and the writer's findings cannot be resolved until both re-examine the outcrops together. However, even if the gravels contain a small percentage of rhyolitic pebbles, this would not necessarily mean that the Tower Creek conglomerate was younger than the rhyolitic welded tuffs in the map area. The tuffs are only the northernmost extension of the welded tuffs that occur in the middle of the thick plateau rhyolites to the south (Boyd, 1961). Rhyolitic pebbles in the Tower Creek conglomerate would suggest that some of the lower rhyolite flows to the south may be older than the conglomerate.

Hague *et al.* (1896, p. 3) reported a vertebrate fossil, identified by O. C. Marsh as the remains of a Pliocene horse, from the Tower Creek conglomerate near Tower Fall. However, no trace of the material exists at the National Museum (C. L. Gazin, personal communication, 1957), and J. T. Gregory (personal communication, 1957) reported that the Peabody Museum of Natural History, Yale University, had no record of the fossil or correspondence between Marsh and Hague mentioning it. The writer believes that the reported fossil is probably valid, for indirect evidence suggests that this group of rocks in the Tower Fall area is probably Pliocene (Fig. 1).

Late canyon basalt. The late canyon basalts (Hague *et al.*, 1896) are petrographically similar to the early canyon basalts and include the thick flows at Overhanging Cliff, Basalt Butte, and Junction Butte. The two most important correlations in the late canyon basalt sequence are between the canyon exposures and (1) the flow capping Junction Butte and (2) the columnar basalts on the south flank of Buffalo Plateau (Fig. 3, sections 3, 4, 5, 13, 14, 15). These outcrops fix the age of the late canyon basalts as post-Lost Creek tuff and pre-Yellowstone tuff.

Yellowstone tuff. Rhyolitic welded tuff of the plateau sequence covers broad areas in the southwest part of the map. Boyd (1961, p. 393) named this unit the Yellowstone tuff. In the map area, the Yellowstone tuff is typically exposed on Mount Everts, 2 miles east of Mammoth Hot Springs, and at Lost Falls, 0.5 miles south of Tower Junction (Pl. 5). Other excellent exposures can be seen in the Yellowstone River canyon south of Tower Fall.

Typical specimens of tuff are gray to lavender, weathering purple to reddish brown. The rock is composed of sanidine and quartz phenocrysts in a vitric groundmass of annealed and flattened shards with rare small oligoclase laths and magnetite inclusions. When the rock is composed chiefly of phenocrysts or is highly pumiceous it is dark gray and resembles basalt in outcrop. Examples of this are south of Oxbow Creek between the west edge of the Crescent Hill basalt and the large area of Yellowstone tuff that surrounds the faulted blocks of Madison limestone (Pl. 1).

The Yellowstone tuff is about 100 feet thick on Mount Everts where it is underlain by a thin black obsidian layer and a 4–8-foot bed of thinly laminated ash. At Lost Falls the tuff is more than 150 feet thick, and large flattened vesicles (1.5 by 0.5 inches) are abundant in the lower 17 feet of the section. Well-developed columnar joints extend from the bottom to the top of the cliff at Lost Falls. Thin sections of tuff taken at 15-foot intervals show no vertical variation in the degree of flattening of annealed shards in the section.

The Yellowstone tuff overlies late canyon basalt and older rocks in the Tower Junction area. It is overlain by basalt flows south of Mount Everts in the west, and it is probably older than local flows in the Lamar River valley (Howard, 1937, p. 74). Fossils have not been reported from the Yellowstone tuff. Indirect evidence suggests that the Yellowstone

tuff in northern Yellowstone Park is possibly late Pliocene. As noted previously, Pliocene vertebrate remains were reported from the Tower Creek conglomerate (Hague *et al.*, 1896). The middle Pliocene Teewinot formation in Jackson Hole contains rhyolitic tuff and is overlain by the Bivouac formation, a succession of conglomerate, rhyolitic welded tuff, and pumicite of uncertain age (Love, 1956a, p. 1911). Scholten *et al.* (1955, p. 373–375) reported rhyolites of middle Oligocene and Pliocene-early Pleistocene age from the Lima region, Montana. Kirkham (1927, p. 37–38) suggested that the Pliocene-Pleistocene rhyolitic "Tertiary late lavas" bordering the Snake River plains of central Idaho were western extensions of the Yellowstone Park plateau rhyolites. The writer therefore suggests that the Yellowstone tuff is at least middle Pliocene and possibly late Pliocene.

Post-Yellowstone tuff gravels and basalts. A fluvial gravel deposit, 20 feet thick, crops out beneath columnar basalt in the Lamar valley between the mouths of Cache and Soda Butte creeks. It contains pebbles of basaltic andesite breccia, basalt, and rhyolitic welded tuff (B. E. Macomber, 1956, M.A. thesis, Northwestern Univ., p. 111) and is possibly younger than the Yellowstone tuff. Basalts overlying the gravel are not displaced by faulting which affected the rhyolitic tuffs. South of Mount Everts extensive basalt flows overlie the Yellowstone tuff.

Basalts of undetermined age. Many local outcrops of basalt cannot be referred to the Tower Fall section. The flows rest on terraces cut in Beartooth Precambrian rocks and early acid breccia along the Yellowstone valley. Other basalts of questionable age crop out above early basic breccia near the summit of Prospect Peak and in the upper reaches of Slough Creek valley.

Glacial and postglacial deposits. Glacial till covers most of the map area but is mapped only where it completely masks the bedrock. Moraines have modified drainage patterns locally, the principal change occurring in the Tower Fall area (Howard, 1937, p. 92–124). Damming of the Yellowstone River by the Lamar Glacier resulted in deposition of lacustrine sands and silts in the canyon lake. Remnants of these are mapped on the east side of the Yellowstone valley opposite Tower Fall.

The glaciated contour of valleys in northern Yellowstone Park is altered locally by landslip, hot spring, and alluvial deposits.

Genesis of the Pyroclastic Rocks

The Cenozoic pyroclastic sequence in Yellowstone Park is cyclic in part. Oldest rocks are dacite to acid andesite in composition followed by local andesitic basalt lavas. Medium andesite pyroclastic rocks grading to basaltic andesite ejecta, followed by local K-rich basaltic lavas, complete the cycle. The cycle repeats again within the late breccia sequence (Hague *et al.*, 1896; 1899). The eruptive sequence represented by the early acid breccia and early basic breccia went from silicic to mafic. The silicic eruptions were apparently more volatile than the mafic outpourings as the early acid breccia contains proportionately more ash than does the early basic breccia. The relative amount of ash in both units decreases up-section.

In light of Kennedy's (1955) concept of magmatic differentiation and distribution of volatiles in a magma chamber, the writer suggests that each eruptive sequence in the Yellowstone Park region was initiated by periodic tapping and drawing off of the upper, volatile-rich, relatively silicic fraction of the source magma. As each cycle progressed, more mafic and less silicic material was ejected. Each cycle terminated with the extrusion of fluid basaltic lavas. Volcanic activity then ceased until further differentiation of the magma produced an upper volatile-rich fraction which later reached the surface and started a new cycle.

The early and late andesitic breccias represent a major period of eruption during early Eocene to early Oligocene time in the Yellowstone region. The rhyolitic floods represent a second major eruptive period during possibly middle to late Pliocene time. Basaltic lavas were extruded intermittently between the two major episodes. In northern Yellowstone Park a limited trachyrhyolitic eruptive phase possibly heralded the later rhyolitic eruptions.

Larsen (1940) suggested that the magma of the Yellowstone-Absaroka petrographic province was derived from a parent basaltic magma through slow differentiation at great depth and that further differentiation at shallower depth, with crystal fractionation and assimilation of granitic material, produced magmas with compositions of the various rock types now exposed. This hypothesis satisfies conditions for the lower Cenozoic andesitic and dacitic rocks and basalts; however, the vast quantity of upper Cenozoic rhyolitic rocks is difficult to account for by this mechanism.

Hamilton (1959) suggested that the plateau

rhyolites were the differentiated upper crust of a Duluth-, Wichita-, or Bushveld-type lopolith that has been forming in the Yellowstone region since early Pliocene time. Hamilton's hypothesis is based on (1) local ophitic olivine basalts associated with the rhyolites, (2) compositional similarity between the Yellowstone rhyolites and "red rocks" of the Duluth, Wichita, and compositional similarity between the Yellowstone rhyolites and "red rocks" associated with lopoliths elsewhere. The analogy that the Yellowstone rhyolites are the differentiated crust of a lopolith is based, therefore, to a large degree on the assumption that the "red rocks" mentioned are really differentiated lopolithic facies. Field relationships of the Bushveld "red

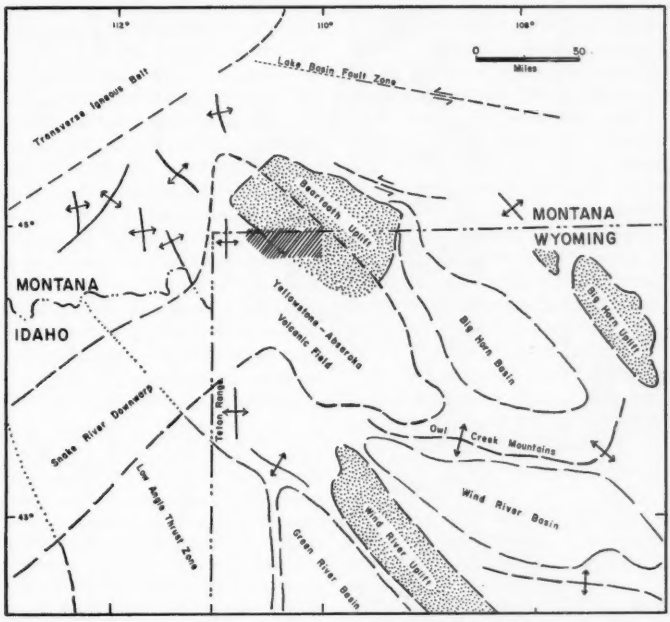

Figure 4. Tectonic map, Yellowstone National Park region. Map area hachured. Data modified after King *et al.* (1949)

Bushveld complexes, and (3) a lack of intermediate rock types in the Yellowstone rhyolite-basalt assemblage. Of prime importance is Hamilton's inference (1959, p. 225) that formation of the Yellowstone structural basin occurred in Pliocene time, prior to eruption of the upper rhyolite flows, rather than as a result of Laramide tectonism followed by cauldron subsidence related to eruption of the Absaroka Mountain sequence. Hague *et al.* (1896, p. 3) suggested that the rhyolites were deposited in a pre-existing structural and topographic basin. Stratigraphic relations between the rhyolitic tuffs and older rocks in the map area lead the writer to agree with Hague.

Hamilton (1959, p. 228) emphasized the rocks" (the tuffs and rhyolites of the Rooiberg series and the Red Granite) indicate that the Bushveld stratiform sheet was intruded, for the most part, between the Rooiberg series and the underlying Pretoria series, that the Red Granite was intruded into the complex and adjacent sedimentary and volcanic rocks after the complex had solidified, and that some of the granophyre associated with the norites in the complex is apparently metamorphosed country rock (Du Toit, 1926, p. 140; van Rooyen, 1951, p. 69; Du Toit, 1953, p. 181, 185–186). Merritt (1958, p. 61) concluded that some of the Oklahoma granite masses in the Wichita Mountain region were batholiths and considerably younger than the gabbros, in

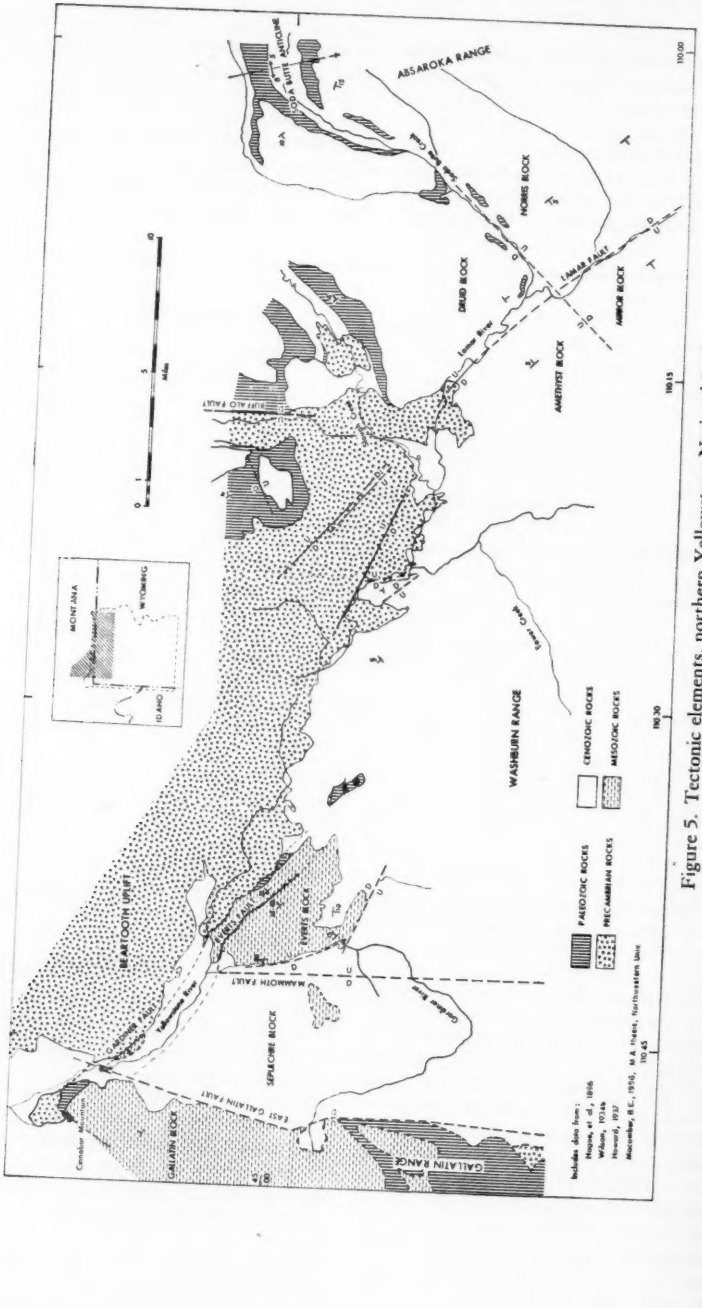

Figure 5. Tectonic elements, northern Yellowstone National Park

which differentiation produced olivine gabbro, troctolite, diallage, gabbro, and anorthosite but not rocks of granitic composition. The "red rocks" associated with the Duluth complex, as described by Grout (1918) and Grout and Longley (1935), intrude the gabbros and may be the products of differentiation, although Grout (1918, p. 636) indicated that some of the Duluth "red rocks" resulted from the assimilation of sedimentary quartzites.

Buddington (1959), in reviewing granite emplacement, discussed many areas which had Cenozoic igneous activity similar to that in Yellowstone Park. Many eruptive rocks, including rhyolitic, trachytic, andesitic, and basaltic associations, in areas that have subsided are related to central parent stocks of granitic material. Structural troughs in New Zealand and northern Sumatra contain extensive rhyolitic tuff deposits, about 1200–2000 cubic miles in volume, similar to the rocks of the Yellowstone basin. According to Buddington (1959, p. 680–681), these rocks formed as the result of breakthrough of shallow magma underlying a pre-existing tectonic depression. Buddington (p. 740) indicated that some small epizonal plutons may be granitic, granophyric, or monzonitic differentiates of basaltic magma, but

"... it seems essential that most of the sialic flows and stocks and all of the batholiths of the epizone originated from granitic magma of deep-seated origin.... If the granite originated as a differentiate of gabbroic magma in the mesozone or catazone, at a level above the source of the gabbroic magma itself, then bodies of gabbro at least 10 times as large as the granite plutons should be common in the mesozone or catazone of the orogens. Such bodies do not occur in appropriate volume in these zones as now exposed."

The present writer suggests that the late Cenozoic rhyolitic magmas of the Yellowstone-Absaroka region were derived by the melting of pre-existing granitic rock at depth. Eruption and emplacement of the Yellowstone tuff is discussed in detail by Boyd (1961). Field work and petrographic examinations by the writer support Boyd's conclusion that the Yellowstone tuff is probably a Katmian sand flow or froth flow deposit that had its source in many eruptive centers throughout the park.

STRUCTURAL GEOLOGY

Regional Structure

Yellowstone National Park is in the Wyoming Rockies on the southwest side of the Beartooth uplift (Fig. 4). Except for the Absaroka Range, a maturely dissected volcanic plateau, the Wyoming Rockies are basically anticlinal flexures modified by normal and reverse faulting. The shelf Rockies on the east are characterized by anticlinal flexures and block uplifts beneath a sedimentary rock veneer 5000–15,000 feet thick. The intensely folded low-angle thrust zone to the west has a 20,000–40,000-foot sedimentary section, and basement rocks are rarely exposed (King *et al.*, 1949; Eardley, 1951; Blackstone, 1956). In southeast Idaho, Rocky Mountain structures are masked by upper Cenozoic volcanic rocks in the Snake River downwarp. Northwest of Yellowstone Park, the Rocky Mountain trend is interrupted by the northeast-trending "Transverse Igneous Belt". The shelf and folded Rockies underwent major deformation during Lancian through Paleocene time. In northern Yellowstone Park, folding and thrusting terminated prior to deposition of the early acid breccia. Following Laramide deformation, epeirogenic uplift of the Beartooth region occurred during Eocene, Oligocene, and Miocene time (Van Houten, 1952). Normal faulting of the Yellowstone tuff suggests that gentle arching of the region continued through Pliocene time. Recent tectonism in the Hebgen Lake area may reflect further regional uplift.

Local Structure

Discussion. Figure 5 indicates the tectonic elements of northern Yellowstone Park. Structural relationships suggest at least five episodes of tectonic activity.

Precambrian metamorphism is indicated by the gneisses and schists beneath the Cambrian Flathead sandstone on Buffalo Plateau.

Laramide movements are indicated by the conglomeratic upper Mesaverde rocks at Mount Everts and overturning of the section adjacent to the Gardiner fault. Crandall(?) conglomerate and related deposits in the region suggest that uplift and stripping of the range cores, accompanied by diminishing deformation, continued to early Eocene time.

Middle Eocene local faulting and fracturing deformed the early acid breccia and Crescent Hill basalt prior to deposition of the early basic breccia.

Late canyon basalts and older rocks were downfaulted in the Yellowstone River valley north of Tower Junction possibly prior to emplacement of upper(?) Pliocene Yellowstone tuff.

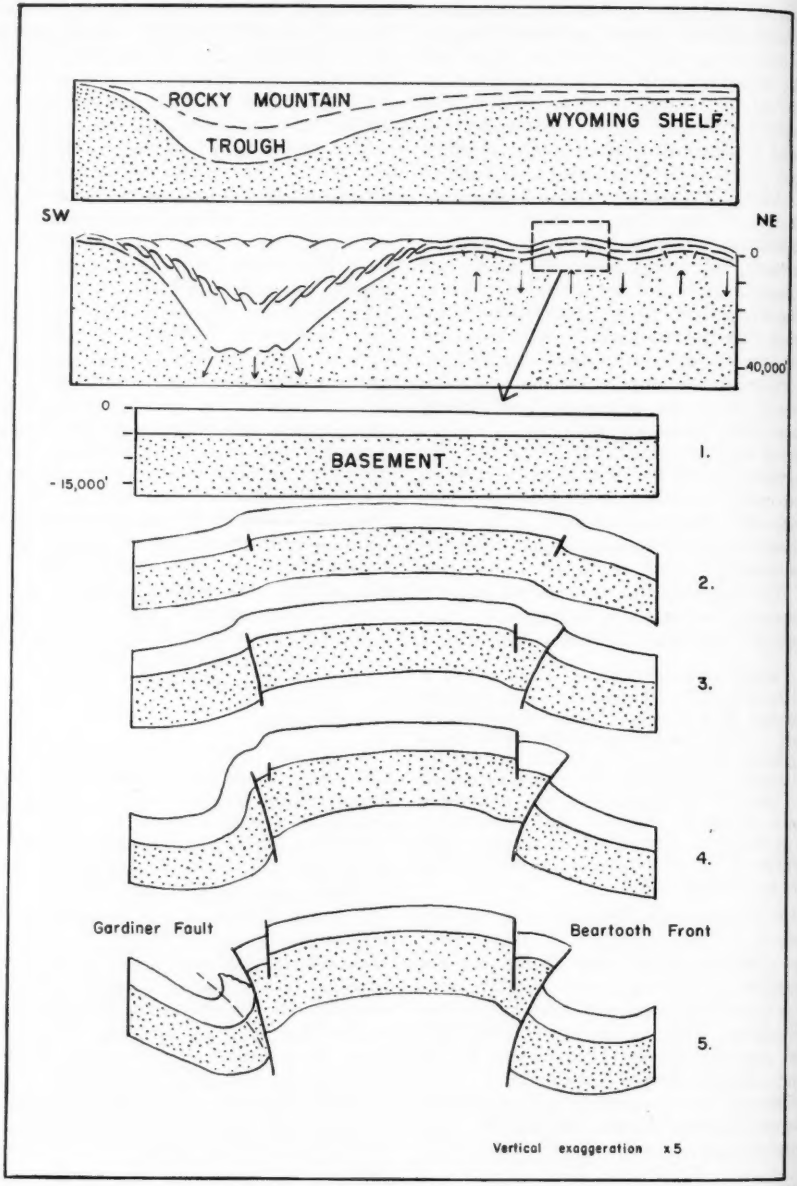

Figure 6. Evolution of the Beartooth uplift

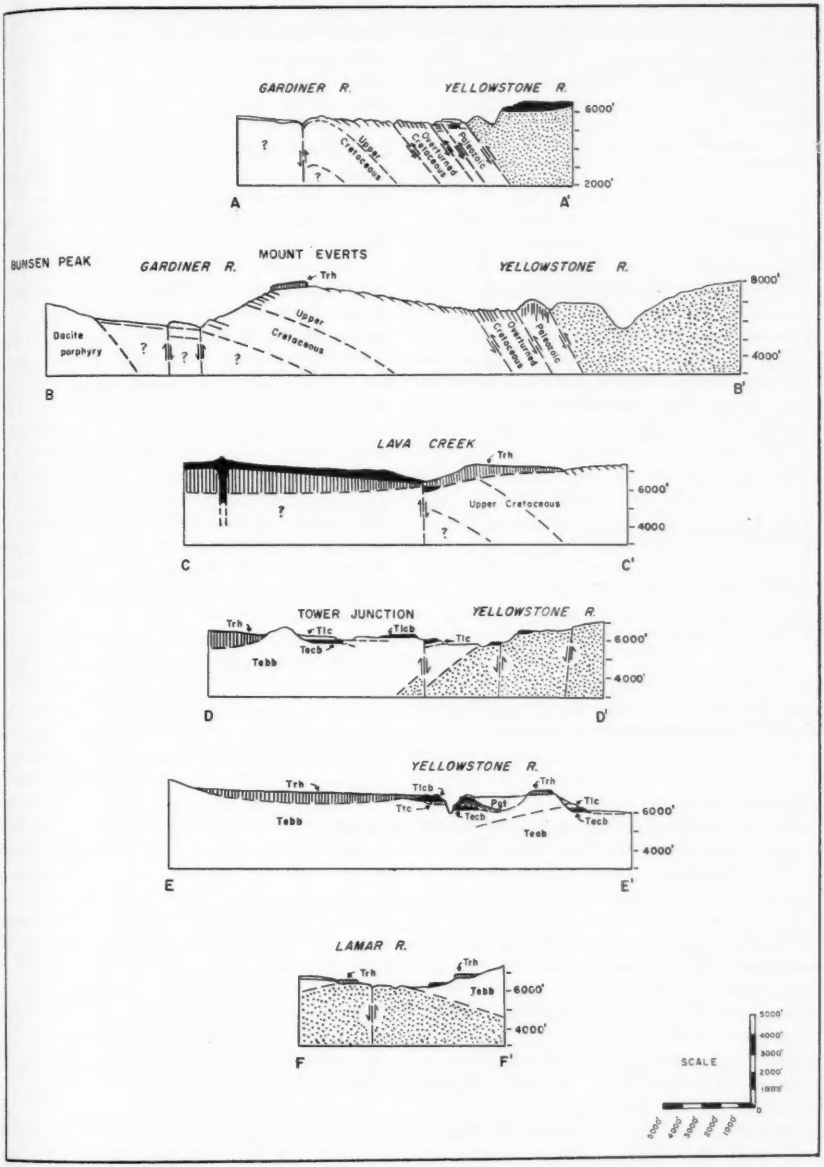

Figure 7. Structure sections, northeast Yellowstone National Park (*See* Pl. 1). No vertical exaggeration of over-all profile, thinner units exaggerated somewhat in thickness for definition: glacial till (Pgt), Yellowstone tuff (Trh), late canyon basalt (Tlcb), Tower Creek conglomerate (Ttc), Lost Creek tuff (Tlc), early canyon basalt (Tecb), early basic breccia (Tebb), and early acid breccia (Teab)

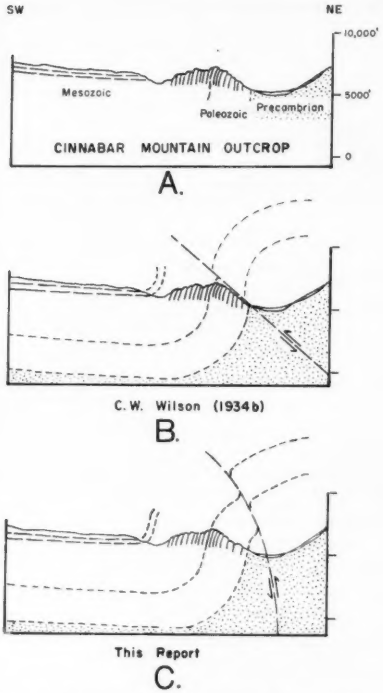

Figure 8. Structural interpretation of the Gardiner fault at Cinnabar Mountain, Montana

Upper(?) Pliocene Yellowstone tuff is displaced 1000–1500 feet (down to the north) along the Lamar fault between Mirror Plateau and The Thunderer. The Yellowstone tuff is also tilted 10° to the south on Mount Everts, and on the south side of the suggested fault in Lava Creek valley the tuff is flat-lying beneath basalt flows. Faults in the plateau rhyolites are common south of the map area (Boyd, 1961).

Gardiner fault system. Hayden (1872), Hague *et al.* (1896), and Bucher *et al.* (1933) reported major faulting in the vicinity of Gardiner, Montana. Wilson (1934b) mapped and named the Gardiner fault (Figs. 6, 7, 8) and suggested that the fault continued southeast beneath the volcanic rocks of Yellowstone Park to join the basement fault on the southwest flank of Rattlesnake Mountain anticline near Cody, Wyoming.

Structure sections across the Gardiner fault (Fig. 7, sections A-A′, B-B′) show imbricate slices of Paleozoic and Mesozoic rocks against overturned upper Cretaceous beds. Imbricate thrusts beneath the main fault plane are exposed along the northwest side of the Yellowstone River, 1.75 miles east of Gardiner, Montana. These imbricate thrusts dip approximately 50°–60° NE., but the attitude of the main fault is not known. The Everts fault is an axial plane fault in the overturned syncline with only a few hundred feet displacement.

Northwest of the town of Gardiner, the Gardiner fault is obscured along most of its trace by basalt, travertine, and alluvium. Four miles northwest of Gardiner, Precambrian rocks are thrust onto overturned upper Cretaceous beds as in the Mount Everts area (Wilson, 1934b, Fig. 3). Three miles farther northwest, at Cinnabar Mountain, structural relationships differ, and the Precambrian rocks are in contact with vertical lower Paleozoic strata which show an incipient flattening of dip to the north at the top of the mountain (Fig. 8a). This change in dip is contrary to the effects of drag folding which should be expected in the footwall block beneath the thrust as shown by Wilson (Fig. 8b). The present writer's interpretation of the fault at Cinnabar Mountain is shown in Figure 8c.

Contrasting structural relationships in the Mount Everts and Cinnabar Mountain areas may be due to steepening of the fault plane toward the northwest or possibly earlier cessation of deformation at Cinnabar Mountain than at Mount Everts. In support of the latter explanation, the writer proposes that during evolution of the Gardiner fault (Fig. 6), deformation of the Cinnabar Mountain and Mount Everts areas was uniform through stages 1 to 4. At this point, displacement along the Gardiner fault was accompanied by left-lateral movement along the East Gallatin fault (Fig. 4). Compression was relieved in the Cinnabar Mountain area as deformation at Mount Everts continued through stage 5. Later normal movement along the East Gallatin fault (Iddings, 1904, p. 104), and development of the Mammoth fault, resulted in downdropping of the Sepulchre block.

Lamar fault and related structures. Howard (1937) recognized the fault origin of the Lamar valley in the southeast part of the map area. Projected dips of the early basalt sheets from Mirror Plateau to The Thunderer and of the Yellowstone tuff from Mirror Plateau to Cache Creek valley suggest that rocks along the northeast side of Lamar valley are downfaulted approximately 1500 feet. Recent investigations

by the writer and B. E. Macomber (1956, M.A. thesis, Northwestern Univ.) lead to the suggestion that a northeast-trending fault in the valleys of Soda Butte Creek and Chalcedony Creek transects the Lamar fault, dividing the area into four blocks (Fig. 5). None of these faults is exposed, but indirect evidence, such as projected dips, juxtaposed unlike pyroclastic facies, straight drainage courses, and dips of individual blocks, suggests their existence. The writer proposes the following partial history of the Lamar valley.

Emplacement of the early basic breccia and early basalt sheets was followed by elevation of the Amethyst block, bringing medium andesite pyroclastic facies into fault contact with basaltic andesite facies of the Druid, Norris, and Mirror blocks. A long period of erosion was followed by emplacement of the Yellowstone tuff. The Druid and Norris blocks were then downfaulted 1500 feet along the Lamar fault. Movement was rotational, in part, with displacement decreasing to the northwest. Differential movement along the fault in Soda Butte Creek valley dropped the Druid block about 200 feet lower than the Norris block, as suggested by elevations and attitudes of Paleozoic limestones near Soda Butte. Movement along the Lamar fault was followed by downward tilting of the Amethyst block to the northeast, displacing Yellowstone tuff 250 feet near Jasper Creek and imparting a northeast dip to the early basic breccia at Fossil Forest.

The inferred fault in Soda Butte and Chalcedony Creek valleys may be part of an important structural element in the Yellowstone Park plateau. Projection of the fault 10 miles southwest aligns it with the Grand Canyon of the Yellowstone River which continues 7 miles farther southwest. Intensive hydrothermal alteration of the plateau rhyolites in the Grand Canyon of the Yellowstone is restricted to a linear zone that may be related to fracturing at depth.

CONCLUSIONS

Volcanic activity in the Yellowstone Park region occurred as an early Cenozoic andesitic phase of considerable volume followed by a late Cenozoic phase of lesser magnitude. The earlier andesitic outpourings contributed to the filling of a Laramide structural basin which may have been further accentuated by cauldron subsidence. A long period of erosion, intermittent eruptive activity, and regional uplift followed during possibly middle and late Oligocene, Miocene, and early Pliocene time. In late(?) Pliocene time, the second period of volcanic activity introduced about 600 cubic miles of rhyolite into the southern two-thirds of the Yellowstone basin. These rocks now constitute the broad north-sloping Yellowstone Park plateau. Faulting of the plateau rhyolites in northeast Yellowstone Park suggests that uplift of the Beartooth region continued through Pliocene time.

Source magmas for the two Cenozoic eruptive phases may have been completely unrelated. The andesitic magma may have been the product of differentiation of a parent basaltic magma with concomitant assimilation of country rock. Later Cenozoic rhyolitic magma was probably derived by melting of pre-existing granitic rock at depth.

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Dispersal Centers of Paleozoic and Later Clastics of the Upper Mississippi Valley and Adjacent Areas

Abstract: Three major dispersal centers of the Paleozoic, Cretaceous, and Tertiary clastics of the upper Mississippi Valley and adjacent areas were located through integrated study of petrography, directional structures, facies maps, and regional stratigraphic relations.

In pre-Mississippian time, nearly all the clastics originally were derived from the Precambrian rocks centered around the Lake Superior region of the Canadian Shield. Relatively modest streams supplied sands and muds and recycling on a stable craton produced mature and supermature sandstones.

The second and third major dispersal centers resulted from Appalachian orogenic activity. Although some contributions from the Lake Superior region and Canadian Shield continued, the tectonic borderlands of the northern Appalachian mountains were the chief dispersal center of Mississippian and Pennsylvanian clastics. In post-Pennsylvanian time, this center shifted to the east of the southern Appalachian mountains. The post-Devonian clastic sediments were transported to the shallow marine shelves, coastal plains, and small deltas of the craton by a series of large, recurring drainage systems.

Despite major shifts in dispersal centers, the slope of the craton in the upper Mississippi Valley and adjacent areas has persisted to the south and southwest throughout Paleozoic, Mesozoic, and Tertiary time.

Inherent in the foregoing regional results are some problems of general interest. These include sedimentary differentiation, intrastratal solution, and the relationships between paleoslope, regional unconformities, and cross-bedding directions.

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INTRODUCTION

In the last decade integrated study of sedimentary rocks has been expanded to include both outcrop and subsurface study and such varied fields as facies mapping, sedimentary petrology, clay mineralogy, and the study of directional sedimentary structures. Although techniques based on all these fields were known and used prior to the last decade, their integrated study has been accelerated in the last few years. This paper reports such an integrated study, emphasizing those techniques useful in a provenance study of cratonic sedimentation.

The objectives of this study were: (1) investigation of the petrology of the Paleozoic, Cretaceous, and Tertiary clastics of the upper Mississippi Valley and adjacent areas; (2) collection and summarization of evidence from directional sedimentary structures, available subsurface facies maps, and stratigraphic relations; (3) integration of petrologic and directional evidence to locate the major dispersal centers of the upper Mississippi Valley and adjacent areas.

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REGIONAL SETTING

The upper Mississippi Valley and adjacent areas lie on the buried Precambrian basement south of the Canadian Shield (Fig. 1). The Appalachian foreland and basin to the east are excluded from this study, and the sample area is thus confined to the craton.

In the area studied the basement is covered chiefly with Paleozoic and lesser amounts of Cretaceous and Tertiary sedimentary rocks that commonly are less than 5000 feet thick

but reach thicknesses of approximately 13,000 and 15,000 feet in the Illinois and Michigan basins, respectively. Four generalized stratigraphic sections of these sedimentary rocks are shown in Figure 2.

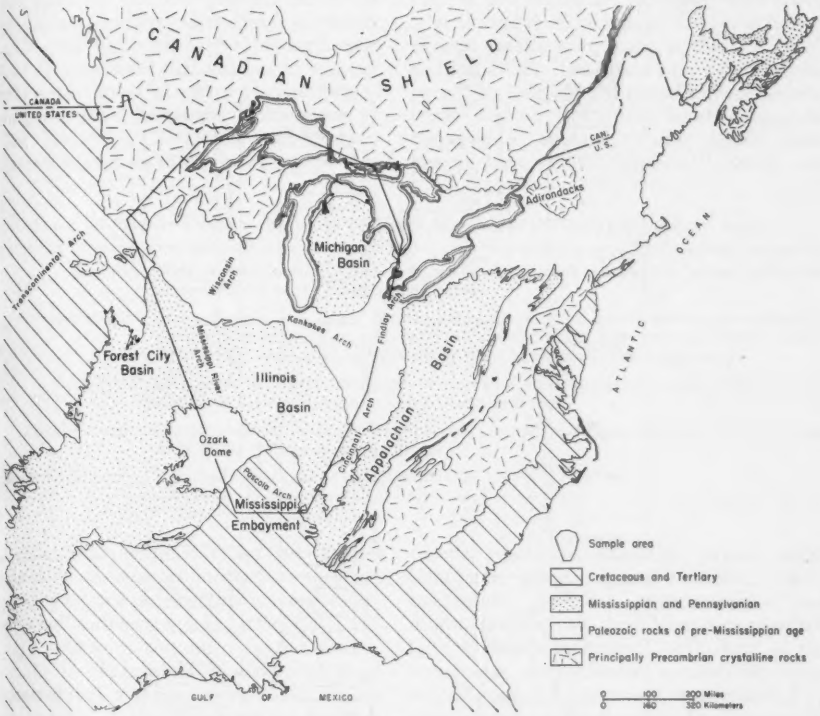


Figure 1. Eastern North America and study area

Data from outcrop observations and available drilling records indicate that throughout much of the sample area the local relief of the Precambrian basement rarely exceeds 500–1000 feet. The sample area contains many classic Precambrian sections of sedimentary rocks, but granite, its allies, and relatively minor metamorphic rocks are the principal basement rocks. In places arkosic sandstones and reddish shale, commonly referred to as the "Red Clastics," occur between sediments of known Cambrian age and the basement complex. The Cambrian sediments generally are considered to be of late Cambrian age.

The last known major tectonic activity in the upper Mississippi Valley region produced a thick sequence of Keweenawan arkoses and interbedded extrusive rocks of late Precambrian age in the Lake Superior region. Since that time the area has behaved as a craton. In Paleozoic time, tectonic activity produced a few broad, regional, recurring arches and structural domes and several intracratonic basins. Early in the Paleozoic, a series of broad regional arches developed—the Cincinnati, Findlay, Kankakee, Wisconsin, and Transcontinental—as well as the Ozark dome (Fig. 1). By the end of Ordovician time the Illinois, Michigan, and Forest City basins were already clearly defined.

The pre-Mississippian sediments of the sample area are of a distinctive type and consist of widespread mature sandstones, associated with carbonates and some evaporites. The individual lithologic units of the pre-Mississippian sequence generally occur as thin, widespread

sheets, which have many minor diastems and unconformities resulting from intermittent exposure to air. Shales are not abundant. In Silurian time little tectonic activity occurred, and even mature quartz sands are absent.

Two major unconformities occur in the pre-Mississippian sequence, one at the base of the Ordovician St. Peter sandstone and the other at the base of the Middle Devonian limestones. Both unconformities truncate older sediments to the north in much of the area and imply that a broad regional rejuvenation of the craton took place northward.

In Mississippian time, widespread carbonates and some evaporites were supplemented by appreciable amounts of shale and less mature siltstone and sandstone. The third major unconformity of the Paleozoic followed Mississippian sedimentation. Like its precursors, this unconformity largely truncates older sedimentary rocks to the northwest, north, and northeast; this again implies a broad regional rejuvenation of the craton to the north.

Pennsylvanian sedimentary rocks are marked by a dominance of shale, immature sandstones, and coal beds. Marine carbonates are present but in minor amounts.

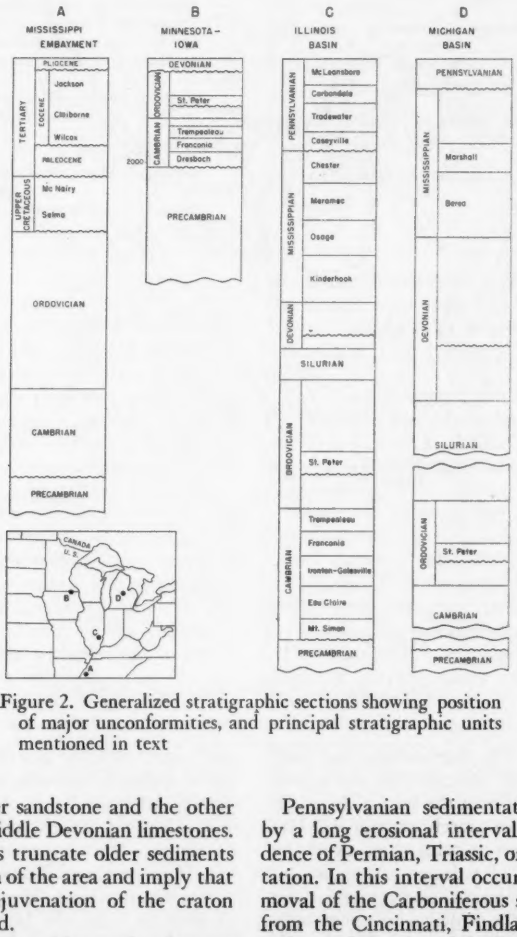


Figure 2. Generalized stratigraphic sections showing position of major unconformities, and principal stratigraphic units mentioned in text

Pennsylvanian sedimentation was followed by a long erosional interval that left no evidence of Permian, Triassic, or Jurassic sedimentation. In this interval occurred the initial removal of the Carboniferous sedimentary rocks from the Cincinnati, Findlay, and Kankakee arches and the rise of the Pascola arch. It produced at the base of the Cretaceous the fourth major unconformity in the area.

In the upper Mississippi Embayment, the basal Cretaceous sediments are of late Cretaceous age and consist of relatively immature, nonmarine clastics and subordinate marine sedi-

ments. Along the northern parts of the Transcontinental arch, the basal Cretaceous is late Cretaceous in age and consists of a mixed sequence of sands, shales, and some carbonates. In the late Cenozoic, broad, regional rejuvenation, probably similar in all but minor details to that which produced the four previous major unconformities, initiated deposition of stream gravels of the Lafayette type along major preglacial drainage lines that were oriented principally southward. Pleistocene glaciers covered most of the upper Mississippi Valley region.

The Paleozoic, Cretaceous, and Tertiary sediments belong to three broad groups. The pre-Mississippian sediments are the typical products of cratonic sedimentation; they are marked by mature and supermature sandstones, widespread carbonates, and little shale. Although carbonates continued to accumulate, especially in the middle Mississippian, large amounts of shale and less mature sandstones are characteristics of the Mississippian, Pennsylvanian, Cretaceous, and Tertiary sediments.

The dominant sedimentary structures of all these clastics, whether carbonates or mature or immature arenites, are those commonly formed in shallow water. Ripple marks and cross-bedding predominate. Turbidites that have graded bedding have never been reported.

Table 1 summarizes the salient features of the study area.

Table 1. Salient Geologic Features of Upper Mississippi Valley and Adjacent Areas

Tectonics	Wide basins separated by broad, regional arches. Gentle deformation (minor folding and normal faulting) of thin sedimentary cover
Unconformities	Many diastems and minor unconformities. Four major pre-Pleistocene unconformities occur at base of St. Peter, mid-Devonian, Pennsylvanian, and Cretaceous. Major unconformities are widespread and commonly truncate older sediments northward
Rock types	Carbonates and mature sandstones diagnostic of pre-Mississippian sections; shale and less mature sandstones diagnostic of post-Devonian section
Lithologic units	Characteristically thin and widely traceable, except for elongate, lenticular Pennsylvanian and Chester (Mississippian) sandstones
Sedimentary structures	Principally ripple marks and cross-bedding

PETROLOGY

Previous Work

Previous petrologic investigations of the Paleozoic, Cretaceous, and Tertiary sandstones of the upper parts of the Mississippi Valley and adjacent regions consist of a large number of local studies and a few more comprehensive regional studies, such as those by Dake (1921), Grim (1936), Thiel (1937), Siever and Potter (1956), Siever (1957), and Pryor (1960). Table 2 summarizes the published studies.

Although one of the earliest detailed studies (Hopkins, 1895) was made on thin sections, most of the studies are concerned with heavy minerals and texture. A large number of heavy-mineral studies were made in the late 1920's and the 1930's. The list of heavy minerals reported is a long one and includes anatase, andalusite, apatite, barite, chloritoid, corundum, dumortierite, epidote, fluorite, garnet, hornblende, hypersthene, ilmenite, kyanite, leucoxene, magnetite, monazite, pyrite, rutile, sillimanite, sphene, spinel, staurolite, topaz, tourmaline, tremolite, zircon, zoisite, and others. Although all these have been reported from Paleozoic sandstones, only three of the nonopaque, nonmicaceous minerals are common. Zircon and tourmaline, supplemented by some garnet, are the principal heavy minerals, except in the basal Cambrian, and commonly form as much as 98 per cent of the entire count. Thus the Paleozoic sandstones have a strikingly mature and stable heavy-mineral suite. In contrast, kyanite, sillimanite, and staurolite are the dominant heavy minerals of the Cretaceous and Tertiary sandstones.

Less study has been devoted to the light minerals. Although several textural studies have been made on light minerals, comparatively few mineralogical investigations of the light fraction have been reported. However, the mature character of most of the pre-Mississippian sandstones is well documented. The less mature Pennsylvanian and Mississippian sandstones have received more petrographic study, especially in the past 10 years. The principal petrographic studies of the Cretaceous and

TABLE 2. PUBLISHED PETROLOGIC STUDIES OF THE SANDSTONES OF THE UPPER MISSISSIPPI VALLEY AND ADJACENT AREAS

Author	Date	Heavy minerals	Petrography	Texture	Age
Allen	1951	X			Eocene
Alty	1933	X			Devonian
Andresen	1961	X	X	X	Pennsylvanian
Andrews	1958	X		X	Cretaceous
Atherton *et al.*	1960	X	X		Pennsylvanian and Chester
Baskin	1956		X		Lower Paleozoic
Becker	1931	X			Cambrian
Berg	1952		X		Cambrian
Biggs and Lamar	1955			X	Carboniferous
Blankenship	1956	X			Tertiary
Clark	1950	X		X	Pennsylvanian
Cordry	1929	X			Cambrian
Crowley and Theil	1940		X		Precambrian and Cambrian
Dake	1918			X	Paleozoic, Cretaceous, and Tertiary
Dake	1921	X		X	Ordovician
Dapples	1955			X	Ordovician
Derry	1934	X			Ordovician
Driscoll	1959	X		X	Cambrian
Ernst	1954	X			Ordovician
Fisk	1951	X		X	Pleistocene
Gault	1939	X		X	Pennsylvanian
Goldich	1934		X		Lower Paleozoic
Graham	1930	X		X	Cambrian
Graham	1933	X		X	Cambrian and Ordovician
Grawe and Cullinson	1931	X		X	Ordovician
Greenberg	1960a	X	X	X	Chester and Pennsylvanian
Greenberg	1960b	X			Chester and Pennsylvanian
Grim	1934	X			Cretaceous
Grim	1936	X	X	X	Paleocene and Eocene
Hamblin	1958	X	X	X	Cambrian
Heald	1956		X		Ordovician
Hopkins	1958	X	X	X	Pennsylvanian
Hopkins	1895		X		Pennsylvanian
Kelly and Beutner	1930	X			Pennsylvanian
Lamar	1928	X		X	Ordovician
Lamar and Grim	1937	X			Cretaceous and Cenozoic
Lamar and Reynolds	1952		X		Pliocene
Lamar and Sutton	1930	X			Cretaceous and Tertiary
McCartney	1931	X			Mississippian
Needham	1934	X			Cretaceous
Ockerman	1930	X			Cambrian
Ostrander	1941	X			Cambrian
Pentland	1931	X			Cambrian
Pinsak	1956	X		X	Tertiary
Potter	1955	X		X	Pliocene
Potter and Glass	1958	X	X		Pennsylvanian
Pryor	1960	X	X	X	Cretaceous
Pryor and Glass	1961		X		Cretaceous and Tertiary
Pye	1944	X	X	X	Mississippian
Rusnak	1957		X	X	Pennsylvanian
Shrode	1954	X			Cretaceous
Shrode and Lamar	1953		X		Cretaceous and Cenozoic
Siever	1949		X		Pennsylvanian

TABLE 2. *Continued*

Author	Date	Heavy minerals	Petrography	Texture	Age
Siever	1953		X		Mississippian
Siever	1957	X	X		Pennsylvanian
Siever and Potter	1956	X	X		Pennsylvanian
Stauffer and Theil	1941	X			Cambrian and Ordovician
Sterns	1933	X			Pennsylvanian
Sun	1954	X			Eocene
Templeton	1951	X			Cambrian
Tester and Atwater	1934		X		Lower Paleozoic
Theil	1935	X		X	Ordovician
Theil	1937	X			Ordovician
Tyler	1936	X			Ordovician
Wayne	1960	X		X	Cretaceous(?)
Wilgus	1933	X			Cambrian
Willman	1942		X		Paleozoic
Willman and Payne	1942			X	Ordovician and Pennsylvanian
Willman and Payne	1943	X		X	Ordovician

Tertiary sediments are those of Grim (1936), Potter (1955), Pryor (1960), and Pryor and Glass (1961).

Objectives and Procedures

Our principal petrologic objective was the characterization of the major mineral associations in an area of predominantly cratonic sedimentation. Figure 3 shows the localities sampled for this study.

The mineral associations were studied principally in thin sections. To estimate the composition of each thin section, 200 point counts per slide were used. The percentage of polycrystalline quartz, defined by sutured and interpenetrating crystals, commonly showing undulose extinction and strain shadows, also was estimated by 200 point counts per thin section. Unusually good examples of polycrystalline quartz grains are shown in Plate 1.

A study of the heavy minerals also was made. Standard petrographic procedures were followed. The size range 0.062–0.500 mm was used rather than a single size fraction, because good hydraulic sorting made choice of a single size fraction difficult. We also wished to appraise the properties of the entire sand fraction rather than merely those of a single grade.

Because of the maturity of most of the Paleozoic sandstones, it was clear that an enumeration of the heavy minerals would contribute little of value. Although each slide was examined for kinds and general abundance of heavy minerals, our principal effort was directed to estimating the percentage of angular tourmaline grains present. Percentage of angular tourmaline provides a quantitative measure of sediment abrasion not easily obtained from thin-section estimates. Grains with roundness of less than 0.6 (Krumbein, 1941, p. 68–70) were classified as angular; grains with roundness of 0.6 or more were classified as round. As used in this paper, "angularity" refers to the proportion of grains with less than 0.6 roundness. Whenever possible, 100 or more tourmaline grains per slide were classified.

The effect of grain size on our estimates of tourmaline angularity and the percentage of polycrystalline quartz present is evaluated by eight fractional analyses (Fig. 4). The contrasts in grain size among the samples are striking. Comparable contrasts of tourmaline angularity have been shown by two quite different techniques (Siever and Potter, 1956, p. 320; Potter and Glass, 1958, p. 29).

But how well does tourmaline angularity correlate with the angularity of the other heavy minerals or of either a single size fraction or the entire size range of the quartz grains? Certainly it would furnish a poor abrasion index if it correlated poorly, no matter how convenient it might be to use.

Appendix 1 describes an experiment specifically designed to answer this question. By use of the multiple correlation coefficient, this experiment demonstrated that tourmaline angularity is highly correlated with the angularity of zircon grains, the angularities of the quartz grains of the .125- to .177-mm size fraction, and the angularity of the quartz grains of the size

fraction 0.062–0.500 mm. Thus tourmaline abrasion is closely identified with the abrasion of the other principal detrital elements, which implies an essentially common abrasion history.

Petrologic data gathered by previous workers were supplemented by additional analyses of thin sections and heavy minerals. Appendix 2 gives these petrographic analyses, polycrystalline quartz abundance, tourmaline angularities, and sample locations. Table 3 summarizes all the available petrographic analyses by system.

In reporting our petrographic analyses we have distinguished between the composition of outcrop and subsurface samples. The outcrop samples almost consistently have less carbonate and more matrix than subsurface samples.

Samples that contain more than 50 per cent carbonate are classified as limestones and are omitted from Table 3.

We have used the petrographic terminology of Pettijohn (1957b, p. 283–339).

Results

Geologic systems. Detailed results are presented for each geologic system represented.

CAMBRIAN: The Cambrian sandstones contain a relatively moderate amount of matrix and very minor amounts of either rock fragments, muscovite, biotite, or chlorite and generally are well rounded and well sorted.

Feldspar content, however, ranges from 0 in 15 per cent of the samples to 40.5 per cent (Appendix 2, sample 339). It averages 8.7 per cent. This estimate might be smaller if more subsurface core samples had been available south of the Wisconsin outcrop belt. As a group, the samples of this study are feldspathic or subarkosic sandstones. One of the characteristics of the Cambrian sandstones is the prevalence of authigenic feldspar (Goldich, 1934; Tester and Atwater, 1934). Authigenic feldspar was noted in approximately 75 per cent of the

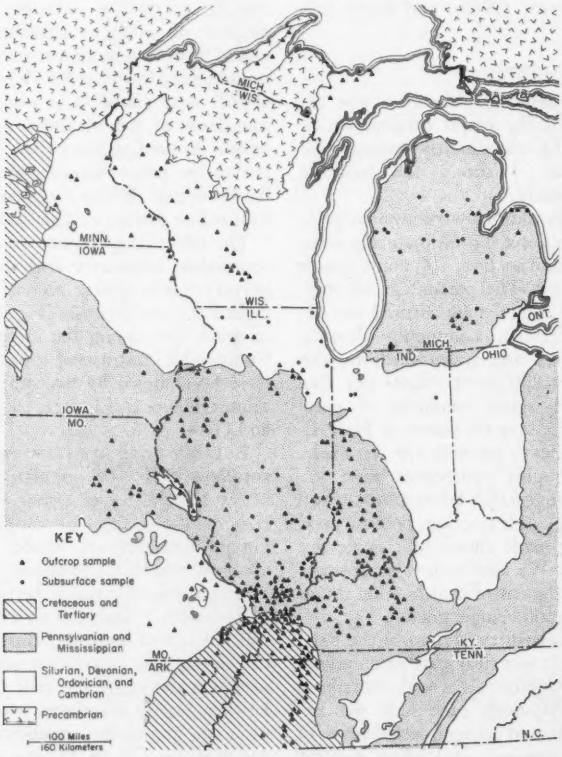


Figure 3. Sample locations in upper Mississippi Valley and adjacent areas

thin sections of this study and was found in all the areas where Cambrian sandstones were studied. Berg (1952, p. 221) reported as much as 48.5 per cent feldspar, much of it authigenic and mostly orthoclase, from the Franconia formation in Minnesota.

Light-yellow-green to dark-green glauconitic oöliths and aggregates also are characteristic of a great many Cambrian sandstones. Glauconite is especially abundant in the sandstones of the Franconia and Trempealeau throughout most of the area. Where glauconite is present, authigenic feldspar generally is well developed. However, authigenic feldspar also occurs where glauconite is absent. Collophane pseudomorphs and fragments of fossil debris also may constitute several per cent of the sample.

Except in some basal Cambrian sandstones, the dominant heavy minerals are stable species and are principally zircon and tourmaline.

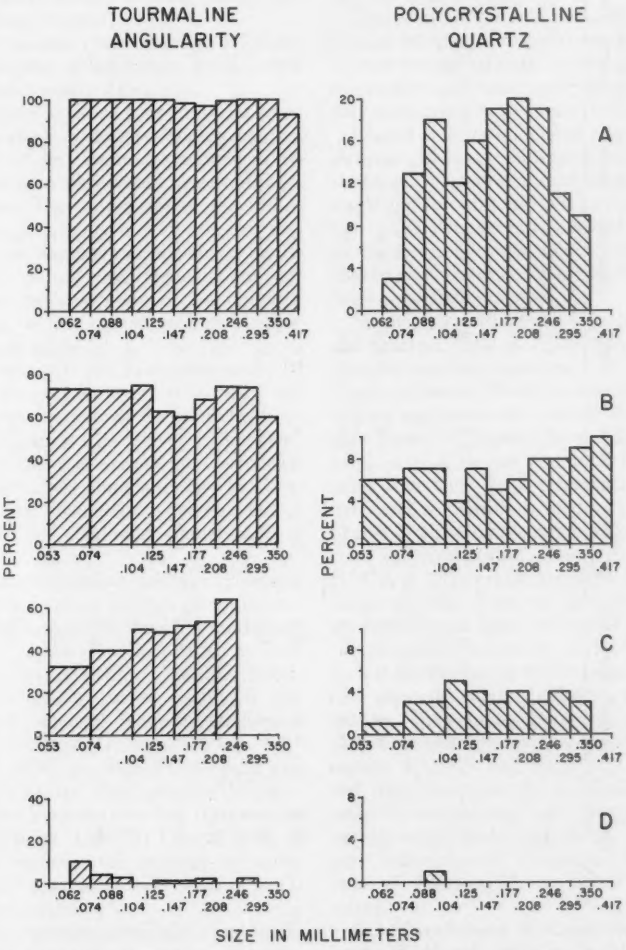


Figure 4. Analyses of tourmaline angularity and polycrystalline quartz fraction in (A) Pennsylvanian subgraywacke, (B) Pennsylvanian protoquartzite, (C) Mississippian protoquartzite, and (D) Ordovician orthoquartzite

TABLE 3. AVERAGE PETROGRAPHIC COMPOSITION OF SANDSTONES OF THE UPPER MISSISSIPPI VALLEY AND ADJACENT AREAS
(Per cent)

	Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Argillaceous rock fragments	Miscellaneous	Polycrystalline quartz	Angular tourmaline	Tourmaline grains counted	Number of analyses
Tertiary												
Outcrop	83.6	9.1	0.0	0.7	2.0	3.7	0.6	0.3	10.5	85.2	903	5
Subsurface	*	..	..	..	..	..	..	..				..
Cretaceous†												
Outcrop	84.0	3.5	0.0	1.5	8.5	2.0	0.0	0.5	8.5	93.2	2397	7
Subsurface†	*	..	..	..	..	..	..	..				..
Pennsylvanian**												
Outcrop	75.3	16.6	2.6	2.6	1.6	0.6	0.9	2.4	7.3	81.1	12,201	161
Subsurface	68.9	14.6	9.1	2.5	1.9	0.4	1.1	1.5				66
Mississippian												
Outcrop	80.0	9.0	6.0	0.4	0.4	0.9	2.3	1.1	4.4	56.4	9220	41
Subsurface	80.7	6.8	8.1	0.8	0.4	0.3	1.8	1.1				39
Devonian												
Outcrop	92.6	0.0	9.0	0.0	0.0	0.0	0.0	0.0	0.3	8.6	848	6
Subsurface	79.0	0.0	20.6	0.0	0.0	0.0	0.0	0.4				8
Ordovician												
Outcrop	95.4	2.6	0.5	0.7	0.1	0.0	0.1	0.6	0.6	9.4	1900	10
Subsurface	89.1	1.4	8.9	0.5	0.0	0.0	0.0	0.1				16
Cambrian												
Outcrop	79.3	5.2	3.5	8.7	0.2	0.6	0.2	2.3	2.6	28.5	3111	40
Subsurface	92.5	2.5	0.0	1.0	0.0	0.5	0.0	3.5				1

*Not determined
†From Pryor (1960)
**Includes data from Siever and Potter (1956, Table 2A); Siever, 1957 (Table 1); Potter and Glass (1958, Table 4)

However, garnet is present in most Cambrian sandstones, and in some formations, as for example in the Jordan, Madison, and Mazomanie (Ockerman, 1930, p. 353; Pentland, 1931, p. 28-33; Ostrander, 1941, p. 146-148; Stauffer and Thiel, 1941, p. 75; and Driscoll, 1959), it is the dominant heavy mineral. It has not been reported in abundance in the sandstones of the Ozark region. Less stable species such as kyanite, monazite, hornblende, staurolite, topaz, sphene, and others are present locally but only in small amounts, except at some places near the Precambrian contact.

Ripple mark and cross-bedding, at places very abundant and of large scale, are the most prominent sedimentary structures.

Although some regional heavy-mineral studies have been made, a detailed, regional petrographic study of the Cambrian sandstone of the upper Mississippi Valley would be rewarding.

ORDOVICIAN: The sandstones of the Ordovician consist principally of well-rounded and well-sorted quartz grains and carbonate. Detrital matrix, micas, and rock fragments are negligible. These sandstones are orthoquartzites. Cross-bedding is not uncommon, and ripple marks are numerous.

Feldspar content is low. In approximately 70 per cent of the samples it is less than 0.5 per cent. In only one sample of the St. Peter from the Michigan basin is it as high as 4.5 per cent. Authigenic feldspar is present in Ordovician sandstones but is less abundant than in Cambrian sandstones. Glauconite is present, but in comparison to the Cambrian sandstones it is rather rare.

Zircon and tourmaline are the principal heavy minerals. In many samples garnet is present as a minor constituent. In a few sandstones such as the Glenwood (Thiel, 1937, p. 118; Willman and Payne, 1943, p. 536-537; Ernst, 1954, p. 1028-1029) garnet locally is the dominant heavy mineral. Some less stable species such as staurolite (Graham, 1933, p. 476-486; Ernst, 1954, p. 1028-1029), pyroxene (Willman and Payne, 1943, p. 536-537), epidote, hornblende, and tremolite (Grawe and Cullinson, 1931, p. 320-326; Derry, 1934, p. 85-87) have been reported, generally in minor abundance and commonly near areas of available crystalline rocks.

Because they consist principally of intensely abraded and mineralogically mature detritus, regional mineral associations are more difficult to establish for Ordovician than for Cambrian sandstones.

SILURIAN: Rocks of the Silurian system in this area are dolomites and limestones and are almost devoid of clastic materials, especially discrete sand bodies. For this reason, only the mineralogy of the clay fraction is considered.

DEVONIAN: The sandstones of the Devonian are supermature orthoquartzites consisting of well-rounded and well-sorted quartz and carbonate. Carbonate is present both as a cementing agent and as locally incorporated fossil debris. Anhydrite is a minor cementing agent in a few samples.

Glauconite is not common, and feldspar and detrital micas are virtually nonexistent. None of these was tallied in this investigation. Zircon, tourmaline, and some rutile are practically the only nonopaque heavy minerals in most slides.

Judged in terms of detrital constituents, thin sections and heavy-mineral slides of Devonian sandstones contain almost nothing but the most stable and resistant end products of sedimentation. Very detailed study would be necessary to reveal regional variations in the petrographic character of Devonian sandstones in the upper Mississippi Valley region.

Although cross-bedding is present (Grabau and Scherzer, 1910, p. 72-75, 81), ripple marks probably are more common.

MISSISSIPPIAN: With few exceptions, Mississippian sandstones are noticeably less mature than those of Ordovician and Devonian age. Both detrital matrix and rock fragments are more abundant. Rock fragments exceed feldspar, and angularity is moderate. As a group (Table 2) these sandstones belong to the lithic clan and correspond closely to Pettijohn's (1957b, p. 291) protoquartzites. However, the values of Table 3 are the average of the three sandstone types listed in Table 4.

Kinderhook sandstones of the Ozark region, such as the Bushberg and Sylamore in Missouri and adjacent parts of Illinois, are the most mature. They consist principally of quartz and carbonate and are notably deficient in mica, matrix, feldspar, and rock fragments. Zircon and tourmaline are the principal heavy minerals. These sandstones are orthoquartzites that are essentially indistinguishable from those in the Ordovician and Devonian. However, this type of Kinderhook sandstone is relatively scarce.

Other sandstones of Kinderhook-Osage age in the Michigan and Illinois basins are of an entirely different petrographic type and are volumetrically important. These siltstones and sandstones—principally the Borden group in

*Not determined
†From Pryor (1960)
**Includes data from Siever and Potter (1956, Table 2A); Siever, 1957 (Table 1); Potter and Glass (1958, Table 4)

TABLE 4. AVERAGE PETROGRAPHIC ANALYSES OF MISSISSIPPIAN SANDSTONES
(Per cent)

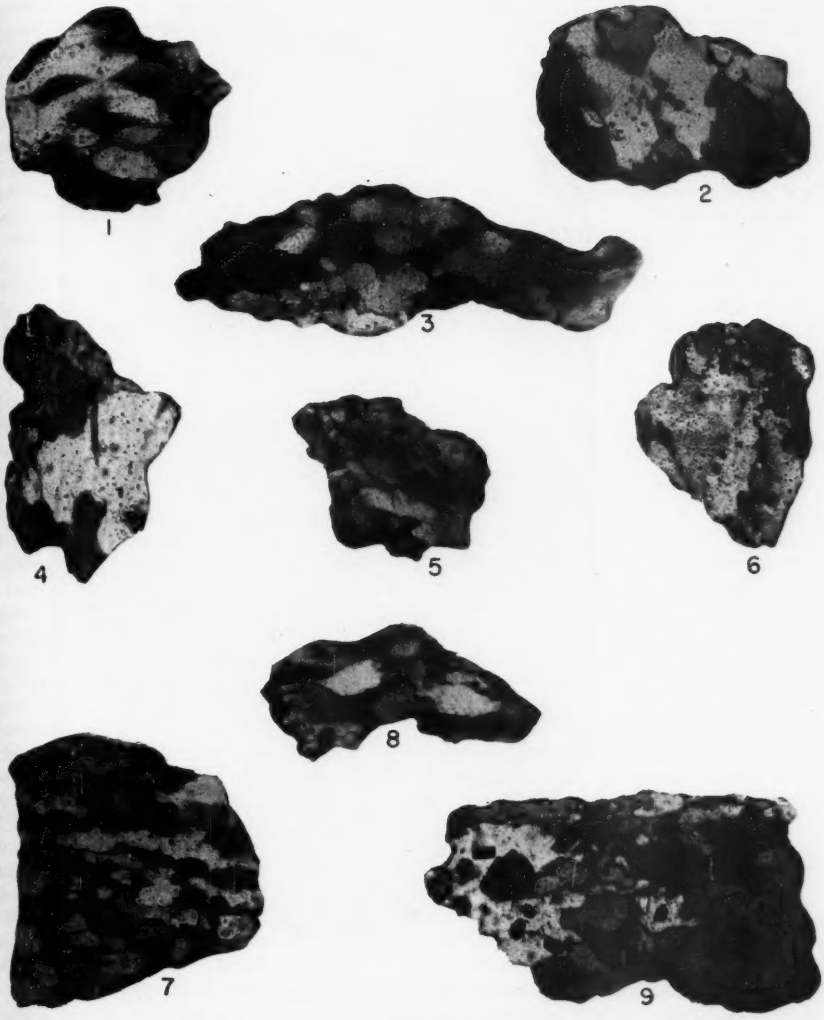
		Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Argillaceous rock fragments	Miscellaneous	Polycrystalline quartz	Angular tourmaline	Number of analyses
Chester and Meramec	Outcrop	81.7	7.5	6.4	0.4	0.2	1.2	1.3	1.3	2.6	55.7	29
	Subsurface	82.0	6.7	8.0	0.7	0.2	0.3	1.3	0.7			33
Osage and Kinderhook*	Outcrop	76.4	13.8	0.6	0.4	0.8	0.4	6.4	1.2	11.9	92.7	6
	Subsurface	72.0	8.1	9.2	1.1	1.8	0.1	4.5	3.2			8
Kinderhook of Ozark region	Outcrop	78.3	2.3	17.5	0.0	0.7	0.0	0.0	1.2	0.3	4.3	3
	Subsurface	..	..	..	..	..	..	..	..	..	..	..

*Exclusive of Kinderhook of Ozark region

the Illinois basin and the Berea formation an Marshall group in Michigan—have abundan matrix, are rich in rock fragments, are angula and contain detrital chlorite and biotite as we as appreciable amounts of muscovite. Neith authigenic feldspar nor glauconite is prominen In most samples, zircon and tourmaline an some garnet are the dominant heavy mineral Stearns (1933, p. 103–106) notes, however, tha in western Michigan the Marshall group sand stones contain minor amounts of hornblende actinolite, and epidote and have more garne than their eastern Michigan equivalents. Petro graphically, all these sandstones are subgray wackes. The sandstones of the Berea formation and Marshall group differ from their equiva lents in the Illinois basin, principally in grain size. In the Illinois basin their equivalents mostly of Osage age, are commonly siltstone rather than sandstones. Although this siltstone sequence is not rich in sedimentary structures some current lineation, ripple marks, and small scale slumping are present. In Michigan th sandstones of the Marshall group have cross bedding and ripple marks (Monnett, 1948, p 669–672).

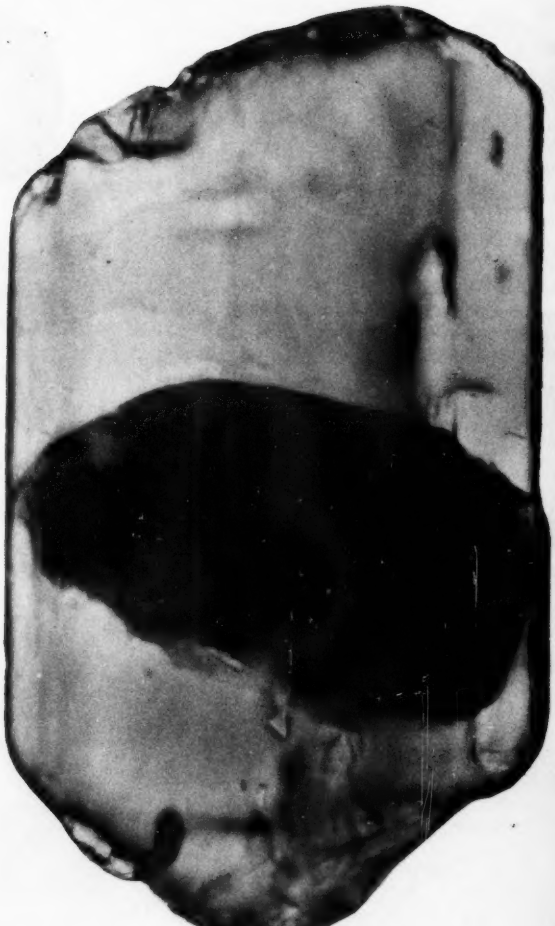
In the Illinois basin these subgraywacke sand stones were followed by important carbonat deposition—the Warsaw, Salem, St. Louis, an Ste. Genevieve limestones—of the Meramec series. The few sandstone lenses in these lime stones and the sandstones of the Chester serie belong to the third type of Mississippian sand stone. This third petrographic type has les matrix, less mica, and fewer rock fragments than the underlying subgraywackes. Argilla ceous rock fragments are more plentiful than feldspar. Authigenic feldspar is not abundant Angularity is moderate. Zircon and tourmalin are the principal heavy minerals, and garne occurs in very minor amounts (McCartney 1931, p. 83–87; Pye, 1944, p. 72–89; Atherto *et al.*, 1960, p. 8).

Oölites and fossil debris commonly are pres ent, reflecting the affinity of the sandstones t marine carbonates. In comparison to the Cam brian sandstones, the quartz sandstones of th Chester have very little glauconite. Carbon aceous matter, macerated into minute shred and fragments, occurs in some samples. Ripple marks and cross-bedding are the most con spicuous sedimentary structures. Load casts convoluted bedding, current lineation, and cut and-fill structures also are present. These sand stones correspond closely to Pettijohn's proto quartzite group.



PHOTOMICROGRAPHS OF POLYCRYSTALLINE QUARTZ GRAINS IN THIN SECTION (× 60–80)

(1) Lafayette gravel (Pliocene), (2) and (3) Eocene sands, (4) McNairy sand (Cretaceous), (5) Mt. Carmel sandstone (Pennsylvanian), (6) Tradewater sandstone (Pennsylvanian), (7) Marshall sandstone (Mississippian), (8) Chester sandstone (Mississippian), (9) Marshall sandstone (Mississippian)



ABRADED TOURMALINE OVERGROWTH ON ABRADED DETRITAL CORE
(× 250) Cretaceous McNairy sand, 1 mile west of Buchanan on State Route 140, Henry County, Tennessee

PENNSYLVANIAN: Pennsylvanian sandstones are appreciably less mature than nearly all earlier Paleozoic sandstones. As a group they have slightly more than 15 per cent detrital matrix, approximately 2 per cent mica, and an equal amount of feldspar. Authigenic feldspar is present but is rather insignificant. There is slightly more feldspar than argillaceous rock fragments. Cementing agents consist of both detrital matrix and carbonate. Siderite serves as a minor cementing agent in many sandstones and in a few is the major cementing agent. Macerated carbonaceous material is present in many samples. All the Pennsylvanian sandstones have the same sedimentary structures: ripple mark and cross-bedding predominate, but current lineation, small-scale slumping, load casts, and cut-and-fill also are present.

Pennsylvanian sandstones belong to four different sandstone types. The most mature of the four types is the basal Pennsylvanian sandstone of western Illinois, the Babylon (Table 5). This sandstone consists principally of quartz, some matrix materials, and very minor amounts of feldspar, chert, and rock fragments. It has the lowest angularity of any Pennsylvanian sandstone in the Illinois or Michigan basins. The basal Babylon sandstone of western Illinois, and possibly a few others which directly overlie lower Paleozoic sandstones, stand somewhat apart from the main sequence of successive Pennsylvanian sandstone types.

This sequence is clearly shown along the present southern edge of the Illinois basin. Here the sandstones of the lowest 400 feet of the Pennsylvanian sequence, the Caseyville group, are mature sandstones that consist principally of quartz and carbonate (in subsurface) with minor amounts of feldspar, mica, and argillaceous rock fragments. Far-traveled, subrounded to well-rounded quartz pebbles are present in minor amounts. Cementing agents consist of an admixture of detrital matrix and carbonate. Angularity is greater than that in the Chester. The principal heavy minerals are zircon and tourmaline, with some sporadic and minor amounts of garnet (Kelly and Beutner, 1930, p. 473; Gault, 1939, p. 131–134; Siever and Potter, 1956, p. 327; Potter and Glass, 1958, p. 27; Greenberg, 1960a, p. 35–44). Some slides contain detrital cahalorite. These sandstones and their lateral equivalents, parts of the Mansfield in Indiana and possibly the Parma in Michigan, are orthoquartzites or borderline protoquartzites depending largely on geographic position.

Sandstones that overlie the Caseyville sequence in southern Illinois are less mature and generally have properties intermediate to those of the basal and higher Pennsylvanian sandstones. Along the southern edge of the Illinois basin this transitional petrographic zone is approximately 500 feet thick. In areas of less rapid subsidence, such as western and northern Illinois and possibly the Michigan basin, the zone is much thinner. The sandstones of this transitional group have more feldspar, matrix, mica, and rock fragments than the Caseyville sandstones. Angularity is appreciably higher. These sandstones clearly belong to the protoquartzite group. The upper boundary of this sandstone type is gradual rather than abrupt.

The sandstones overlying the transitional zone form a very homogeneous and widespread series. These sandstones are rich in clay matrix, average 11.5 per cent feldspar, and have appreciable amounts of mica. Chlorite and biotite supplement muscovite. Argillaceous rock fragments are commonly less abundant than feldspar, and glauconite is very rare. Pyrite is abundant in many thin sections and heavy-mineral slides. Angularity is unusually high. Although zircon and tourmaline remain dominant, garnet becomes important (Potter and Glass, 1958, Table 5). These sandstones correspond best, although not precisely, to the subgraywacke group of Pettijohn (1957b, p. 316–321). These higher Pennsylvanian sandstones are nearly indistinguishable from those of Marshall group of the Michigan basin and, except for grain size, from the Borden group of the Illinois basin.

With some modification, dependent on overlap and amount of basin subsidence, this immature sequence is characteristic of most of the higher Pennsylvanian sandstones of the upper Mississippi Valley and adjacent areas. Detailed study probably would reveal regional variations similar to those noted by Siever and Potter (1956) in the basal Pennsylvanian sandstones.

CRETACEOUS: The upper Cretaceous sandstones in the Mississippi Embayment consist of angular, well-sorted quartz grains. Feldspar content is uniformly low, ranging only up to 3.5 per cent. Authigenic feldspar is essentially absent. High percentages of muscovite mica are characteristic of the Cretaceous sandstones (Pryor, 1960), ranging from 1.5 to 10.5 per cent. Biotite and chlorite are rare, as are rock fragments. The matrix is composed chiefly of clay and quartz silt.

The dominant heavy minerals are less mature

TABLE 5. AVERAGE PETROGRAPHIC ANALYSES OF PENNSYLVANIAN SANDSTONES
(Per cent)

		Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Argillaceous rock fragments	Miscellaneous	Polycrystalline quartz	Angular tourmaline	Number of analyses
McLeansboro, Carbondale, Upper Tradewater	Outcrop	58.2	19.0	6.3	7.2	3.4	0.5	3.4	2.0	High	98.8	10
	Subsurface	55.4	16.1	16.8	3.3	3.6	0.3	2.1	2.2			9
Lower and Middle Tradewater*	Outcrop	..	..	..	..	..	..	..	..	Moderate to high	83.6	..
	Subsurface	73.4	8.5	11.7	2.2	1.0	0.4	1.9	1.0			10
Babylon†	Outcrop	92.8	6.3	0.0	0.2	..	0.4	0.2	..	Very low	40.5	6
	Subsurface	..	..	..	..	..	..	..	..			..
Caseyville†	Outcrop	..	..	..	..	..	..	..	..	Moderate	64.8	7
	Subsurface	79.4	2.4	16.0	0.9	0.0	0.6	0.7	0.1			

*These are the "Transitional Group" of Potter and Glass (1958, Table 4).
†From Siever and Potter (1956, Table 2A)

species and are principally kyanite, sillimanite, and staurolite. Tourmaline and zircon occur in small but significant amounts. The Cretaceous heavy-mineral assemblage of the embayment region differs from that of the Eastern Gulf Coast assemblage (Needham, 1934) in its low epidote content. One of the rare accessory minerals is graphite (Shrode, 1954; Pryor, 1960). Although Cretaceous sandstones have good sorting and little feldspar, the combined abundance of mica, matrix, chert, and argillaceous rock fragments indicates that they are protoquartzites (Pettijohn, 1957b, p. 95–300).

Cross-bedding is a ubiquitous, prominent sedimentary structure in the Cretaceous sandstones. However, ripple mark, small-scale slumping, and cut-and-fill also are present.

These sandstones grade southward into argillaceous and carbonate sediments (Pryor, 1960).

TERTIARY: The sands of the Tertiary closely resemble those of the Cretaceous and occur as two distinct types. The Eocene sands are well sorted, angular, and fine-grained (Grim, 1936, p. 102), whereas the Pliocene sediments commonly consist of coarse sands and gravels (Potter, 1955, p. 7). As in the Cretaceous sands, argillaceous rock fragments and feldspar are not abundant. Matrix, mica, and chert together constitute nearly 15 per cent of the sands. The matrix is chiefly clay and quartz silt. There is noticeably less muscovite mica and more chert in Eocene than in Cretaceous sands. Eocene sands are protoquartzites. The heavy-mineral assemblages of the Eocene and Pliocene sandstones are similar to that of the Cretaceous but have somewhat less kyanite.

Cross-bedding and cut-and-fill structures are abundantly present in both Eocene and Pliocene sands.

Tourmaline and polycrystalline quartz. Tourmaline angularity was assessed by examining 30,980 grains.

In Cambrian sandstones tourmaline angularity averages 28.5 per cent. It is appreciably less in Ordovician and Devonian sandstones, in which it is 9.4 and 8.6 per cent, respectively. In Mississippian sandstones, tourmaline angularity averages 56.4 per cent. The Pennsylvanian, Cretaceous, and Tertiary sandstones average, respectively, 81.1, 93.2, and 85.2 per cent angular tourmaline grains. Thus the Cambrian, Ordovician, and Devonian sandstones are dominantly well rounded, the sandstones of Mississippian age have average moderate rounding, and those of Pennsylvanian, Cretaceous, and Tertiary ages are poorly rounded, as measured by tourmaline angularity. Figure 5 shows examples of the angularity of the tourmaline grains from these three groups.

Abundance of polycrystalline quartz was determined by a count of approximately 80,000 grains.

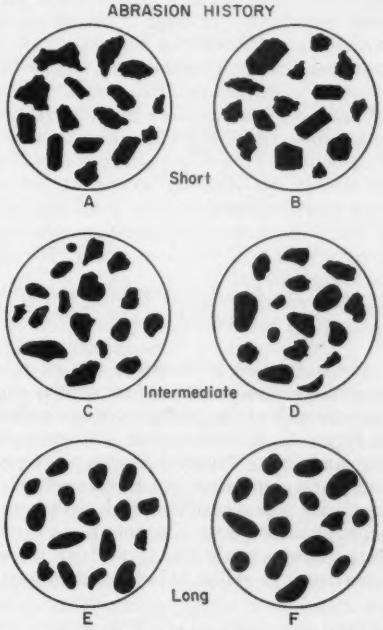


Figure 5. Some typical tourmaline angularities shown in composite field of view: (A) Cretaceous protoquartzite, (B) Pennsylvanian subgraywacke, (C) and (D) Mississippian protoquartzites, (E) Devonian orthoquartzite, and (F) Ordovician orthoquartzite

Polycrystalline quartz averages 2.6 per cent in Cambrian sandstones. In Ordovician and Devonian sandstones, it averages only 0.6 and 0.3 per cent. Many Devonian and Ordovician thin sections are essentially devoid of polycrystalline grains. The average value for Mississippian sandstones, however, is 4.4 per cent. As much as 22 per cent was noted in some slides from sandstone of the Marshall group of Michigan. Polycrystalline quartz averages 7.3, 8.5, and 10.5 per cent in Pennsylvanian, Cretaceous, and Tertiary sandstones. Some thin sections of Pennsylvanian sandstone contain as much as 22 per cent polycrystalline quartz (Siever, 1957, Table 1).

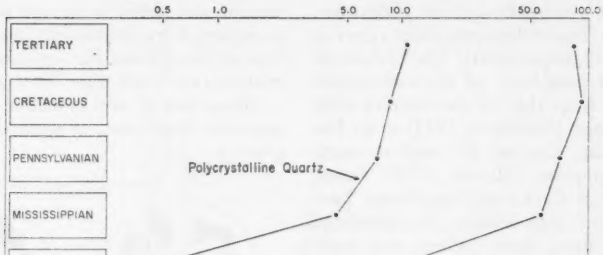


Figure 6. Average values of tourmaline angularity and polycrystalline quartz by system. Note correlation.

The correlation between the average values of polycrystalline quartz and tourmaline angularity in each of the geologic systems is shown in Figure 6. In each system low tourmaline angularity is accompanied by low polycrystalline quartz abundance, and a high incidence of angularity is paralleled by a high abundance of polycrystalline quartz. This correlation extends to the various sandstone types within the several systems as well as to individual samples.

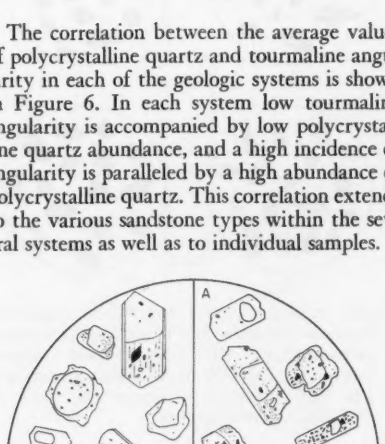


Figure 7. Composite field of view showing overgrowths on tourmaline cores in (A) Tertiary-Cretaceous protoquartzites, (B) Pennsylvanian subgraywackes, and (C) Mississippian protoquartzites. Commonly both core and overgrowth are abraded.

The tourmaline grains exhibit another significant property. A very small percentage, perhaps 1 grain in 5000, of the Mississippian, Pennsylvanian, Cretaceous, and Tertiary tourmaline grains shows clearly defined overgrowths. Although tourmaline overgrowths have been noted in the pre-Mississippian sandstones of this region (Alty, 1933), they are exceedingly rare. Rounded overgrowths on rounded to subrounded detrital cores are the significant feature of these grains. Figure 7 illustrates such grains, and Plate 2 shows an unusually fine example.

Clay minerals. The clay minerals of the associated shales also exhibit broad groupings. The predominant clay mineral in pre-Pennsylvanian shales is illite. Mixed-lattice materials follow illite in abundance in Mississippian, Devonian, Silurian, and Ordovician shales. In the Pennsylvanian, illite continues to predominate, but chlorite and kaolinite rather than mixed-lattice materials are next in abundance. Cretaceous and Tertiary shales have a strikingly different clay-mineral suite in which montmorillonite and kaolinite are the dominant clay minerals (Pryor and Glass, 1961). In contrast to Paleozoic shales, the kaolinite of Tertiary and Cretaceous shales is poorly crystallized, and illite is a minor rather than major constituent.

Interpretation

The significant variations in the petrographic properties of the upper Mississippi

Valley area clastics are tourmaline angularity, abundance of polycrystalline quartz, abundance and kinds of micaceous minerals, amount of feldspar, and the relative dominance of heavy and clay minerals. Table 6 summarizes these properties, dividing them into three broad groups: pre-Mississippian, Mississippian-Pennsylvanian, and post-Pennsylvanian. These three groups reflect rather closely the major tectonic events of much of eastern North America since Precambrian time.

The low angularity, a good sorting, deficiency of argillaceous rock fragments and micas, and general absence of matrix indicate that, as a group, the pre-Mississippian sandstones have had an intense abrasion history that generally led to mature and supermature sandstones.

The extent to which crystalline rocks contributed directly to pre-Mississippian sandstones varies greatly. Although well rounded and well sorted, many of the Cambrian sandstones contain abundant detrital feldspar and garnet. Moreover, hornblende, topaz, kyanite, staurolite, and other minerals of moderate to low stability are not entirely restricted to the basal Cambrian sandstones but are found also in some younger Cambrian sandstones. Such facts suggest that Precambrian crystalline rocks were contributors to many Cambrian sandstones.

Although Ordovician sandstones such as the New Richmond contain some feldspar, the striking predominance of quartz, zircon, and tourmaline in Ordovician sandstones points to a source predominantly consisting of sediments, most of which were already relatively mature. The unstable heavy minerals that are present in minor amounts have a restricted occurrence and are found in regions proximal to exposed crystalline rock areas, such as the southern edges of the Canadian Shield and the Ozark dome. The abundance of garnet and presence of staurolite in the Glenwood sandstone in portions of Minnesota and Wisconsin is especially noteworthy.

In the Devonian, first-cycle sandstones derived from crystalline rocks were almost nonexistent. The extreme mineralogical and textural maturity of the Devonian sandstones indicates a source that contained only mature sandstones.

With the exception of sandstones such as the Babylon, Bushburg, and Sylamore, all volumetrically insignificant and all principally derived from lower Paleozoic sandstones, Mississippian and Pennsylvanian sandstones are less mature.

The high angularity and abundant argillaceous rock fragments and micas of the Kinderhook and Osage subgraywackes in the Illinois and Michigan basins prove that these sandstones could not have been derived from underlying lower Paleozoic sandstones. These sandstones, therefore, reflect an entirely new source in which mature sandstones were relatively scarce. The large volumes of these sandstones make it improbable that the major source region lay within the sample area. The abundant detrital matrix and micas imply poor sorting and hence probably rapid sedimentation at the site of deposition. In both the Illinois and Michigan basins most of these sandstones represent marine deposition (Monnett, 1948, p. 682–688; Swann *et al.*, 1953). With the exception of the western part of the Michigan basin, the heavy minerals in these subgraywackes do not differ greatly from those of the pre-Mississippian sandstones.

The protoquartzites of the Chester and Caseyville sandstones have properties intermediate to those of the mature pre-Mississippian sandstone and the Mississippian and Pennsylvanian subgraywackes. The Chester and Caseyville sandstones represent detritus derived from the mature pre-existing sandstones such as the lower Paleozoic sandstones, from less mature sandstones such as the Mississippian subgraywackes, and possibly from some crystalline rocks. The large volume of sediments suggests that much of this material came from beyond the sample area. The broad petrologic homogeneity of the Chester sandstones in the Illinois basin indicates that the great bulk of this detritus was well mixed by the time it reached the basin. The basal Pennsylvanian sandstones, however, show a western and eastern mineral association (Siever and Potter, 1956, p. 320–322). The greater immaturity of the detritus in the eastern association indicates a resurgence of tectonic activity.

The younger Pennsylvanian subgraywacke sandstones dominantly reflect this now major tectonic activity. Mature pre-existing sandstones made almost no contributions to these sandstones. Immature sandstones were probably the chief source and were supplemented by some low-rank metamorphic and crystalline rocks. Both the large volume and wide area covered by Pennsylvanian sediments indicate that the immature sandstones of this sequence very largely represent detritus introduced to

TABLE 6. SUMMARY OF PETROGRAPHIC VARIATION IN CLASTICS OF THE UPPER MISSISSIPPI VALLEY AND ADJACENT AREAS

System	Angularity	Polycrystalline quartz	Micaceous minerals (sand fraction)	Feldspar	Dominant heavy minerals	Clay-mineral associations*
Tertiary						
Cretaceous	High to very high	High	Abundant, consisting of muscovite and some biotite and chlorite	Low	Kyanite, sillimanite, and staurolite	Montmorillonite ≧ kaolinite ≫ illite > mixed lattice
Pennsylvanian				Moderate	(Minor garnet)	Illite≫chlorite ≧ kaolinite ≧ mixed lattice
Mississippian	Moderate	Moderate to high		Low		Illite≫mixed lattice > chlorite > kaolinite
Devonian	Very low	Virtually nonexistent	Rare; muscovite	Virtually nonexistent	Zircon and tourmaline	Illite≫mixed lattice > chlorite ≧ kaolinite
Ordovician						
Cambrian	Low	Low	Glauconite plus some muscovite, chlorite, and biotite	Moderate; authigenic feldspar very common	(Appreciable garnet)†	Illite≫kaolinite > chlorite ≧ mixed lattice

*Ranked in order of abundance from left to right
†Plus hornblende, epidote, kyanite, and staurolite near Precambrian contact

the region. Relatively rapid sedimentation and subsidence probably account for the poor sorting at the site of deposition.

The tourmaline overgrowths provide a revealing insight into the behavior of this source area. Abraded overgrowths on abraded cores imply authigenesis or low-rank metamorphism of relatively mature lower Paleozoic sandstones. Erosion in Mississippian and Pennsylvanian time released these grains and transported them to the region west of the Cincinnati and Findlay arches, partially rounding the overgrowths. Thus, the dominant Mississippian and Pennsylvanian source area had already had at least one episode of orogenic activity in the lower Paleozoic.

The Cretaceous and Tertiary protoquartzite sandstones of the Mississippi Embayment constitute the third major suite. Judged only by angularity, muscovite mica, feldspar, and argillaceous rock fragments, these sandstones could have had a source broadly similar to that of the Mississippian and Pennsylvanian subgraywackes. Their heavy-mineral suite precludes this possibility, however. The principal heavy minerals of these sandstones are kyanite, staurolite, and sillimanite; this indicates that the source of the Cretaceous and Tertiary sandstones, unlike that of the preceding Paleozoic sandstones, was dominated by high-rank metamorphic rocks such as mica schists and gneisses. The abrupt change in source rock from Paleozoic to Cretaceous time indicates major rejuvenation and uplift of a new source region during the erosional interval between the Pennsylvanian and Cretaceous and thus reflects the first major contribution of detritus from a major metamorphic terrain located outside the sample area. As in Mississippian and Pennsylvanian sandstones, some tourmaline grains with subrounded overgrowths on abraded detrital cores are present and imply rejuvenation and metamorphism of pre-existing sandstones in the source region. The small quantities of feldspar associated with these less mature heavy minerals of the Cretaceous and Tertiary sandstones are best explained by deep weathering of the regolith of the source region. The rarity of well-rounded grains indicates that preexisting sandstones with appreciable abrasion history had no significance as source rocks. Parts of the Tertiary sandstones were, however, derived from the underlying Cretaceous.

Considered in broad perspective, the pre-Mississippian, Mississippian-Pennsylvanian, and post-Pennsylvanian sandstones can be given a relatively simple petrographic interpretation. The progressive increase in maturity of the Lower Paleozoic sandstones can be considered as the result of progressive attrition in a sedimentation system to which crystalline rock contributions were only important in Cambrian time. Progressive reworking on stable marine shelves produced the petrographic properties of this sequence. Although mature sandstones of the pre-Mississippian type contributed detritus, especially to Chester and Caseyville sandstones, the immature detritus of the Mississippian and Pennsylvanian sandstones represents the initial contribution of a major new source to the region west of the Cincinnati and Findlay arches. Immature pre-existing sediments, low-rank metamorphic rocks, and quartzofeldspathic terrains were the principal source rocks. In the Cretaceous and Tertiary a distant region containing high-rank metamorphic rocks and yielding immature sandstones was the principal source of supply.

The clay minerals, like the heavy minerals, illustrate major differences between the Paleozoic sediments and those of the Cretaceous and Tertiary of the Mississippi Embayment.

Where illite is the dominant clay mineral in the Paleozoic sedimentary rocks (Grim *et al.*, 1957; Weaver, 1959), montmorillonite and/or kaolinite (depending upon the facies) are the dominant clay minerals in the Cretaceous-Tertiary sediments. Kaolinite constitutes the chief minor clay mineral in the Cambrian, Mississippian, and Pennsylvanian sediments, whereas chlorite or mixed-lattice materials are the chief minor constituents in the Ordovician, Silurian, and Devonian sediments.

Murray (1953, p. 60), Keller (1956, p. 2704–2708), Degens *et al.* (1957, p. 2443), and Weaver (1959, p. 166; 1960) have demonstrated that kaolinite is a common constituent of nonmarine and littoral Paleozoic sedimentary rocks, whereas illite is common to the marine. Pryor and Glass (1961) and Groot and Glass (1960) also have shown that kaolinite is dominant in the fluviatile and littoral argillaceous materials and that montmorillonite is dominant in the marine shelf sequence of the Cretaceous and Tertiary of both the Mississippi Embayment and Atlantic Coastal Plain. These relationships suggest a strong environmental control of clay mineralogy for both the Paleozoic and Cretaceous-Tertiary argillaceous sediments.

The alteration of montmorillonite to a mica type of mineral (illite) may account for its general absence in sediments older than Meso-

TABLE 7. CROSS-BEDDING AND PRESENT STREAM PATTERN OF THE UPPER MISSISSIPPI VALLEY AND ADJACENT AREAS

Unit	Cross-bedding distribution									Total	Mean	Age	Area	Author
	1 to 40	41 to 80	81 to 120	121 to 160	161 to 200	201 to 240	241 to 280	281 to 320	321 to 360					
Present stream pattern (1)	48	26	47	91	163*	94	50	41	84	644	188	Recent	Upper Mississippi Valley	This paper
Lafayette gravel (2)	10	4	4	16	22	19	27	35*	21	158	264	Pliocene	West Kentucky, southern Illinois, and south Missouri	Potter (1905)
Wilcox (3)	2	2	6	3	14*	13	3	1	2	46	199	Eocene	West Kentucky, southern Illinois, and southeast Missouri	This paper
McNairy (4)	14	37	29	62	140	236*	206	134	62	920	236	Cretaceous	Arkansas, west Kentucky, south Illinois, southeast Missouri, and west Tennessee	Pryor (1960)
Tradewater, Carbondale, and McLeansboro (5)	23	20	11	40	70	95*	68	58	41	426	232	Middle and upper Pennsylvanian	West Kentucky and Illinois	Potter and Glass (1958); this paper
Caseyville, Mansfield, and Parma (6)	26	42	27	91	156	209*	152	65	24	792	214	Dominantly lower Pennsylvanian	West Kentucky, Illinois, Indiana, and Michigan	Potter and Siever (1956)
Chester (7)	81	98	94	155	199	298*	202	127	93	1347	215	Upper Mississippian	West Kentucky, Illinois, and Indiana	Potter *et al.* (1958)
Marshall (8)	0	0	1	0	2	19*	6	1	1	30	228	Early middle Mississippian	Michigan	Briggs (Personal report)
Glenwood-St. Peter (9)	11	8	11	18	21	22	22	19	8	140	220	Ordovician	Illinois	This paper
New Richmond (10)	0	4	6	7	6	7	2	0	0	32	155	Ordovician	Illinois	This paper
Trempealeau (11)	2	1	8	2	8	14*	5	1	4	45	199	Cambrian	Southwestern Wisconsin	Farkas (1960)
Franconia (12)	68	136	186	210	207	280*	214	144	66	1511	192	Cambrian	Minnesota and Wisconsin	Farkas (1960) and Hamblin (1958)
Miners Castle (13)	11	12	7	11	57	106*	102	51	31	388	244	Cambrian	Upper Michigan and adjacent Ontario	Hamblin (1958)
Dresbach (14)	12	13	35	48	72*	54	41	12	10	297	184	Cambrian	Minnesota and Wisconsin	Farkas (1960) and Hamblin (1958)
Chapel Rock (15)	15	5	5	0	3	8	11	42	69*	158	325	Cambrian	Upper Michigan and adjacent Ontario	Hamblin (1958)
Jacobsville (16)	92	27	1	0	0	0	3	27	135*	285	356	Cambrian	Upper Michigan and adjacent Ontario	Hamblin (1958)

Millers Castle (13)	11	12	7	11	57	106*	102	51	31	388	244	Cambrian	Ontario	Hamblin (1958)
Dresbach (14)	12	13	35	48	72*	54	41	12	10	297	184	Cambrian	Minnesota and Wisconsin	Farkas (1960) and Hamblin (1958)
Chapel Rock (15)	15	5	5	0	3	8	11	42	69*	158	325	Cambrian	Upper Michigan and adjacent Ontario	Hamblin (1958)
Jacobsville (16)	92	27	1	0	0	0	3	27	135*	285	356	Cambrian	Upper Michigan and adjacent Ontario	Hamblin (1958)
Barron (17)	0	0	0	0	0	2	2	0	0	4	240	Keweenawan	Northwestern Wisconsin	Sievер (*in* Pettijohn, 1957b)
Bayfield (18)	74	98*	36	4	1	1	0	0	28	241	46	Keweenawan	Northwestern Wisconsin	
Copper Harbor (19)	25	13	13	4	10	10	10	18	29*	132	350	Keweenawan	Upper Michigan	W. S. White (Personal report)
Freda (20)	44	12	5	1	5	22	79	97*	60	325	317	Keweenawan	Upper Michigan	W. S. White (Personal report)
Nonsuch (21)	4	4	0	1	2	12*	8	4	3	38	259	Keweenawan	Upper Michigan	W. S. White (Personal report)
Baraboo (22)	14	20	38	75*	51	45	27	6	7	283	160	Pre-Keweenawan	Southwestern Wisconsin	Brett (1955)
Lorrain (23)	0	7	27	40*	35	29	7	4	2	151	162	Pre-Keweenawan	North shore of Lake Huron	Pettijohn (1957a)
Sioux (24)	2	5	7	7	4	6	5	2	1	39	158	Pre-Keweenawan	Southeast South Dakota, northwest Iowa, and southwest Minnesota	Brett (*in* Pettijohn, 1957a)
Menard (25)	0	3	8*	1	0	0	0	0	0	12	93	Pre-Keweenawan	Upper Michigan	Pettijohn (1957a)
Mississagi (26)	104	192	263	315*	190	72	43	32	49	1260	118	Pre-Keweenawan	North shore of Lake Huron	McDowell (1957)
Sturgeon (27)	0	3	21	30*	23	2	2	0	0	81	134	Pre-Keweenawan	Upper Michigan	Trow (*in* Pettijohn, 1957a)
Total cross-bedding (28)	634	766	849	1141	1298	1580*	1247	880	746	9141	194	Pre-Keweenawan through Pliocene	Greater upper Mississippi Valley region	

*Modal class interval

zoic (Grim, 1953, p. 364) and suggests that postdepositional diagenesis is the most important factor in the differences in clay mineralogy between the Paleozoic and Cretaceous-Tertiary sediments. Differential preservation also is a contributing factor as is evidenced by the fact that nonmarine and littoral facies are less commonly preserved in older than in younger sediments (Gilluly, 1949, p. 580). Hence, there is less likelihood of a kaolinite-rich shale in older sediments.

dispersal centers from which the detrital sediments of the sample area came.

Although others are present, cross-bedding is the most useful sedimentary structure for determining transport direction. The cross-bedding in all these sandstones is similar. Both planar and trough cross-bedding (McKee and Weir, 1953, p. 387) are present. Typically, cross-bedded units are less than 30 inches thick, although units as much as 10 feet thick have been noted.

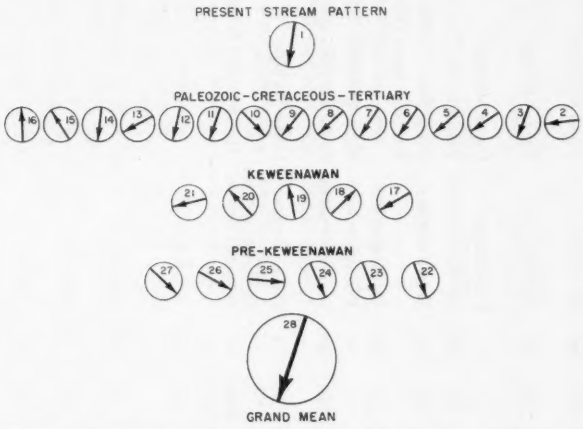


Figure 8. Mean values of present stream pattern and cross-bedding of pre-Pleistocene sandstones. Numbers refer to units of Table 7.

Weaver (1958; 1959; 1960) suggests that clay minerals in sedimentary rocks strongly reflect the character of their source materials and are only slightly modified by their depositional and postdepositional environments. However, Keller (1956) cautions against the use of clay minerals alone as indicators of source-area environments and suggests that depositional environment has an equally important role in the clay-mineral assemblage of sedimentary rocks. Pryor and Glass (1961, p. 51) list provenance, depositional environment, and diagenesis as the most important variables that control the final clay-mineral assemblages in sedimentary rocks. The clays, therefore, are of only relatively minor value in identifying the major dispersal centers.

DIRECTIONAL EVIDENCE

General

Information derived from studies of cross-bedding, major unconformities, and facies maps provides the major key to location of the major

In the last 10, and especially in the last 5, years, considerable cross-bedding data have been gathered in the upper Mississippi Valley and adjacent areas. Table 6 summarizes this information, both published and unpublished. Altogether, 9141 observations were obtained on formations ranging in age from Pliocene through pre-Keweenawan[1]. Because of the difficulties of correlation, the Precambrian units of Table 7 are divided into two groups, Keweenawan (late Precambrian) and pre-Keweenawan (middle Precambrian).

Table 7 also includes the orientation of the present stream pattern in the region. Appendix 3 gives the details of our sampling procedure.

Mean values were computed for each unit of Table 7 by using the Tukey test (Harrison, 1957, p. 98–105; Rusnak, 1957, p. 53–54). This

[1] Since June 1959, several thousand additional cross-bedding observations have been made in Paleozoic sandstones of the study area. These have not been included in Table 7; they confirm the implications of the table.

test yields vectorial means independent of a choice of origin. All the means are significant at the 0.95 level, and most are significant well beyond the 0.995 level. The grand mean of all 9141 cross-bedding observations is 194° (azimuth). The probability that this distribution could have arisen from the sampling of a uniform distribution of cross-bedding directions is 2.48×10^{-154}, which is almost nil.

The mean values of Table 7 fall into three rather clearly defined stratigraphic groups: pre-Keweenawan (middle Precambrian), Keweenawan (late Precambrian), and post-Precambrian. Figure 8 shows the distribution of these mean values. Table 7 and Figure 8 indicate that the dominant cross-bedding direction of the pre-Keweenawan sandstones is southeastward. In contrast, the cross-bedding means of the Keweenawan sandstones are more disperse but suggestive of a dominant northwestward direction. With the exception of the Chapel Rock and Jacobsville sandstones of upper Michigan, almost all the other younger Paleozoic, Cretaceous, and Tertiary sandstones have southwestward-oriented cross-bedding. The vectorial mean of the present stream pattern of the sample area, 188°, is significant far beyond the 0.05 level and differs by only a few degrees from the grand mean, 194°, of all the pre-Pleistocene cross-bedding of the same area.

Essential to interpretation of a paleoslope is the reliability of the information on cross-bedding and major unconformities and of the facies maps on which estimates of paleoslope are based. In our study, the paleoslope interpretation was developed by beginning with the most recent sediments and working backward in time.

Recent and Pleistocene

The present regional slope of most of the sample area is to the south. The orientation of the higher-order streams in the area closely reflects this regional slope, the mean value of which is 188°.

Although we did not make comparable studies of the widespread Pleistocene drift sheets, the general direction of sediment transport in both glacial till and outwash deposits was commonly between 135° and 225° and hence varies but little from the present slope. The direction of preglacial drainage, as depicted by Horberg (1950, p. 98) and Thornbury (1958, p. 451), is essentially the same as that of the present streams.

Tertiary and Cretaceous

In the Pliocene the transport direction of the fluvial Lafayette-type gravels on the Smithland (Bentley) and Lancaster-Ozark

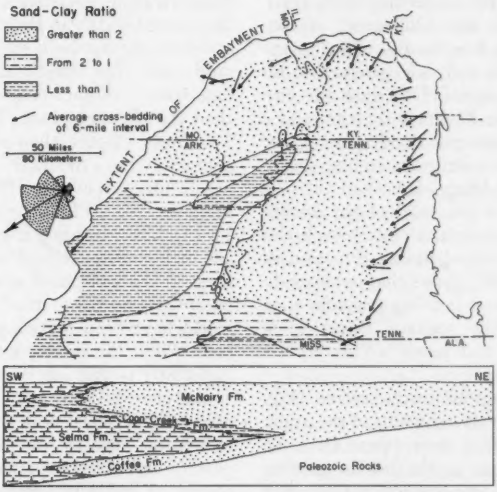


Figure 9. Lithofacies, cross-bedding, and cross section of McNairy formation in upper Mississippi Embayment (adapted from Pryor, 1960, Figs. 2, 13, 17)

(Williana) erosion surfaces of the upper Mississippi Embayment and adjacent areas also closely parallels the present stream pattern (Potter, 1955, p. 125–127). Thus, excluding contributions from the ancestral Tennessee River in parts of western Tennessee and Kentucky, the drainage and sediment-transport pattern in Pliocene time was dominantly to the south and southwest over much of the southern parts of our study area.

The Eocene sediments truncate the underlying Cretaceous and form a series of three units (Wilcox, Claiborne, and Jackson) offlapping southward. The succeeding later Tertiary units (Miocene and Oligocene) unconformably overlie the Eocene and offlap to the south. Cross sections and paleogeographic reconstructions by Stearns (1958) show the progressive offlap of the Eocene sediments and a gradation from nonmarine conditions in the north to nonmarine conditions to the southwest. The cross-bedding directions in the Eocene sandstones at the head of the Mississippi Embayment harmonize closely with this data. The average cross-bedding direction of the Eocene sands, 199°, shows that in this sedimentation system cross-bedding provides a reliable measure of the regional direction of sediment transport and hence accurately reflects the paleoslope.

The lithofacies map and cross section (Fig. 9) of the late Cretaceous sediments in the upper Mississippi Embayment show coarse fluviatile and deltaic sediments in the north, grading southwestward into fine-grained marine-shelf clastics and chalks. The mean value of the cross-bedding of the McNairy sandstone, 236°, indicates that the dominant direction of sediment transport was from the northeast toward a marine-shelf environment to the southwest. Thus, as in the Eocene sands, late Cretaceous regional cross-bedding agrees closely with the major distribution of marine, littoral, and nonmarine facies and accurately reflects the paleoslope.

Pennsylvanian and Mississippian

The Pennsylvanian and Mississippian sandstones of the Illinois and Michigan basins have cross-bedding directions very similar to those of the Cretaceous and Tertiary sands of the Mississippi Embayment. Average values for the Caseyville (and equivalents) and post-Caseyville sandstones are 214° and 232°. The average cross-bedding direction for the Chester sandstones of Mississippian age in the Illinois basin is 215°, and in Michigan the Mississippian Marshall group sandstone has an average value of 228°.

Interpretation of these data is significant in determining the regional slope and is based on (1) knowledge of the environment of deposition and (2) the internal consistency of cross-bedding direction with petrologic evidence and patterns of truncation along major unconformities.

Evidence of the environment of deposition is closely related to the distinction between sheet and channel phases of Mississippian and Pennsylvanian sandstones. This distinction has been noted widely in Pennsylvanian sandstones and recently has been emphasized for Chester sandstones. The sheet phase has a transitional base and is thinner-bedded and finer-grained than the channel phase. The channel phase is 3–6 times thicker than the sheet phase and commonly has a distinctly lenticular, meandering outline (Mueller and Wanless, 1957, Fig. 7; Hopkins, 1958, Pl. 1; Potter *et al.*, 1958, Figs. 11–14). Cross-bedding is best developed in the channel sandstones and subparallels the long axes of channel-phase sand bodies. Although exceptions do exist, most channel-phase sandstones have been regarded in the literature as, principally, the products of fluvial deposition. According to this interpretation, the Pennsylvanian cross-bedding and Chester cross-bedding reflect the regional slope of Carboniferous sedimentation surfaces just as the present stream pattern reflects the present regional slope. A combined low-lying coastal plain and marginal shallow marine shelf appears to provide the best large-scale model of the environment of deposition.

Other evidence supports this interpretation of the paleoslope. Swann *et al.* (1953, p. 1480–1481) have shown the existence of a paleoslope trending southward in the subgraywacke Osage siltstones of the Illinois basin. Pinsak (1957, p. 44–48) has shown the same slope for the Salem limestone in Indiana. These interpretations agree with those of Sterns (1933, p. 111) and Monnett (1948, p. 682) who emphasized a dominant eastern and a secondary western source for the Marshall group sandstones of Michigan. Cohee and Underwood (1944) have shown that the Berea formation has a northeast trend in eastern Michigan. Moreover, the Marshall cross-bedding of eastern Michigan also has a southwestward orientation that, though younger, strikingly parallels the subsurface Berea trend (Fig. 10).

Major carbonate deposition followed this early immature siltstone and sandstone and was in turn succeeded by the alternating limestone, shale, and sandstone sequence of the Chester series. Figure 11 shows Chester cross-bedding, an isopach of a portion of the interbedded limestones, and the major trend of Chester sand deposition. These relations show that the main sandstone trends with their southwestward-oriented cross-bedding extended into areas more favorable for marine carbonate deposition, indicating a dominant transport direction down the paleoslope.

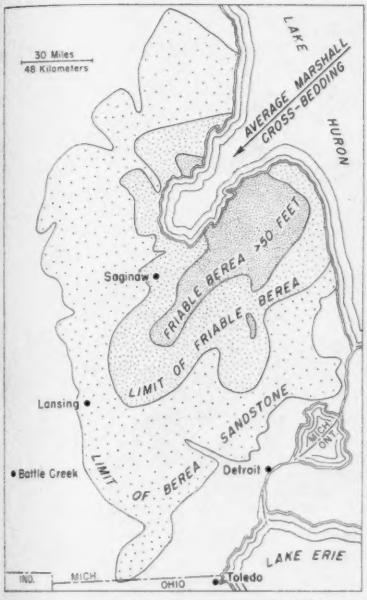


Figure 10. Berea formation and Marshall formation cross-bedding of Michigan basin (adapted from Cohee and Underwood, 1944, Fig. 1)

With the termination of Chester sedimentation in the Illinois basin, this regional slope to the southwest continued to prevail. In the Illinois basin the Mississippian-Pennsylvanian unconformity truncates older rocks to the northwest, north, and northeast, indicating a generally southward direction of tilting. Even in an area as negative as the Michigan basin, erosion was sufficient to remove any evidence of post-Meramecian sediments. Moreover, the stream-cut channels of the Mississippian-Pennsylvanian unconformity that have been mapped in subsurface by Siever (1951, p. 561) and Wanless (1955, p. 1764–1765; Fig. 2) show a dendritic pattern of drainage to the southwest. The good correspondence between this system of fluvially entrenched channels to both the underlying Chester and overlying Pennsylvanian cross-bedding demonstrates the reliability of cross-bedding for inferring paleoslopes in these sediments. Therefore, at least as early as Osage time, the Mississippian and Pennsylvanian clastics of the Illinois basin accumulated on a paleoslope dipping dominantly southwest. This paleoslope prevailed throughout important periods of marine carbonate deposition, was the controlling factor in both the pattern of regional truncation and the orientation of the entrenched channels of the Mississippian-Pennsylvanian unconformity, and continued in the Pennsylvanian. Although evidence is less comprehensive, the Pennsylvanian sediments of the Michigan basin appear to have had a paleoslope similar to that of the Illinois basin in Pennsylvanian time. The same is true for the Mississippian sediments of the Michigan basin, although restricted basin conditions implying negligible paleoslope existed part of the time.

Lower Paleozoic

Exposure and sandstone development restrict most studies of cross-bedding of the lower Paleozoic sandstones to the Ordovician and Cambrian sandstones of Wisconsin, Minnesota, and upper Michigan. Except for the Jacobsville and Chapel Rock sandstones of upper Michigan, all these sandstones have southerly and southwesterly cross-bedding directions.

Our evidence of the direction of sedimentation agrees with the opinions of previous workers. Dake (1921, p. 207–210), Lamar (1928, p. 59–62), Thiel (1935, p. 610–611), and Tyler (1936, p. 82) inferred a northern source for the Ordovician sandstones. The mean values of the Glenwood-St. Peter and New Richmond cross-bedding, 220° and 155°, support these conclusions. Dake (1921, p. 212–216), DuBois (1945, p. 29–33), and Dapples (1955, p. 445–446) also emphasized the northward transgressive character of the St. Peter sandstone. Although recognizing an ultimate northern source, Dapples (1955, p. 464–467) considered the St. Peter sandstone to represent a largely littoral deposit whose main transport direction parallels the strand line.

Petrographic studies of the Cambrian sandstones of Wisconsin, Minnesota, and Illinois

have indicated a dominant northern source (Graham, 1930, p. 715; Pentland, 1931, p. 34; Templeton, 1951, p. 158; Driscoll, 1959, Fig. 5). Stratigraphic studies in southeastern Minnesota and southern Wisconsin (Twenhofel *et al.*, 1935, p. 1714–1715; Berg, 1954, p. 876–877) indicate that the Cambrian sediments form a transgressive-regressive sequence with deposi-

Petrologic, isopach, and structural data also imply a paleoslope interpretation for the northwest-oriented cross-bedding of the Jacobsville and Chapel Rock sandstones of upper Michigan (Hamblin, 1958, p. 51–57, 93–95).

The pre-Middle Devonian and pre-St. Peter unconformities of the upper Mississippi Valley region support a paleoslope interpretation of

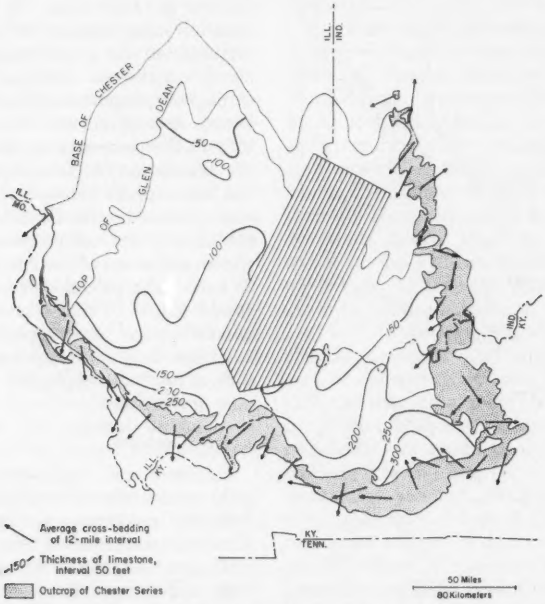


Figure 11. Chester cross-bedding, limestone isopach (Glen Dean to base of Aux Vases sandstone) and principal sand trends. Arrow diagrammatically indicates preferred location and trend of Chester sandstones. Adapted from Potter *et al.* (1958, Figs. 15, 17)

tional strike approximately northwest. By integrating stratigraphic relationships, size data, and regional heavy-minerals studies, Driscoll (1959, Fig. 5) has shown how this transgressive-regressive concept extends from southwestern Wisconsin to northern Michigan. This implies a paleoslope, in dominantly marine sediments, to the south and southwest. The available cross-bedding data broadly correspond to these directions. Hamblin (1961) extended the synthesis of Driscoll and supplemented it with cross-bedding data. Figure 12 shows Hamblin's interpretation and illustrates both the broad south-southwestern direction in most of the area as well as the northwestern direction in upper Michigan.

Cambrian and Ordovician cross-bedding. Both unconformities are best developed along basin margins and proximal to the region's mildly positive areas, such as the Transcontinental arch, the Wisconsin arch, and the Ozark dome. In northeastern Illinois the pre-St. Peter relief exceeds 250 feet (Suter *et al.*, 1959, Fig. 226). Excluding areas around the Ozark dome and some of the regional arches, both unconformities generally truncate older sediments to the north toward the flanks of the Canadian Shield. Figure 13 shows examples of these relations. In both instances greater truncation along the margins of the Canadian Shield show regional rejuvenation to the north with resultant paleoslopes to the south, southwest, and southeast.

This also is reflected in an isopach of the pre-St. Peter Paleozoic sediments of the upper Mississippi Valley and Michigan basins (Cohee, 1948, Fig. 2; Swann *et al.*, 1951, Fig. 153). This isopach, the result of both regional thickening to the south, introduction of new units in the southern part of the area, and pre-St. Peter erosion, shows marked thinning northward toward the Canadian Shield.

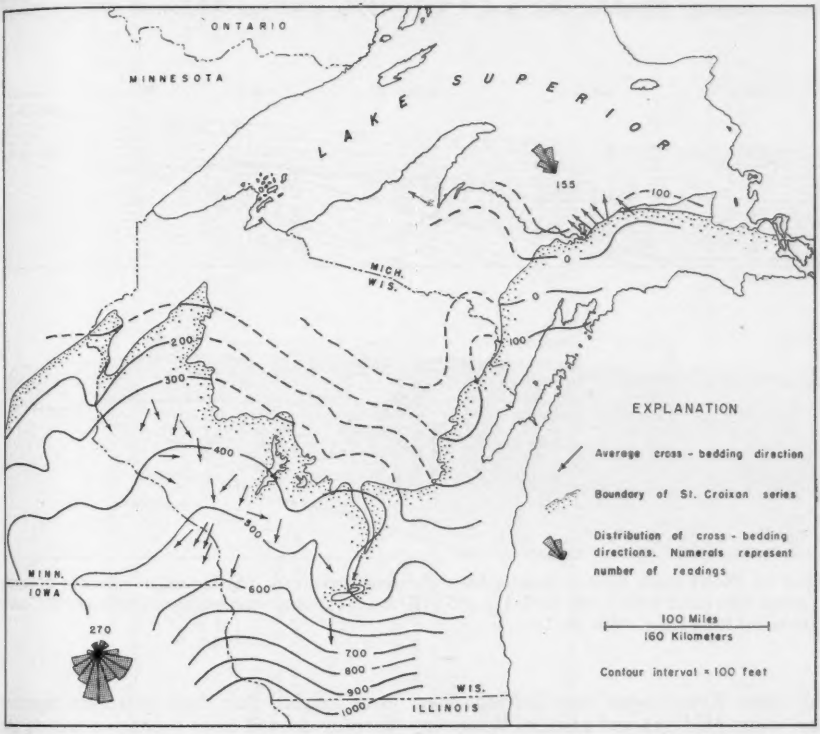


Figure 12. Isopach and cross-bedding directions of Dresbach formation (from Hamblin, 1961, Fig. 5)

Although more data are needed, the relationships between cross-bedding, unconformities, and facies relations of the lower Paleozoic sediments are in general similar to those of the Mississippian and younger sediments. These similarities point to a dominant paleoslope interpretation of the cross-bedding orientation for most of the Ordovician and Cambrian sandstones, many of which have a marine-shelf origin.

Thus, throughout most of the 500-million-year post-Precambrian history of the upper Mississippi Valley and adjacent areas, the paleoslope, which dips gently south-southwest, has remained almost unchanged. Closed basins of essentially zero paleoslope that developed in negative areas and a few local reversals to the north are the only important exceptions. The Devonian and Silurian evaporites of the Michigan basin and less extensive Mississippian evaporites in the Illinois basin indicate the existence of such closed basins. Perhaps the best example of a northeastward paleoslope is provided by the Devonian deposits of the Michigan basin (Briggs, 1959, Figs. 2, 7). The stratigraphic and areal scope of our data indicates, however, that exceptions are insignificant compared to the persistence throughout Paleozoic and later time of the south-southwesterly paleoslope.

Precambrian

The cross-bedding directions of the Pre-

cambrian sandstones of the sample area also contribute to interpretation of the Paleozoic-Cretaceous-Tertiary sequence. The red clastics (Jacobsville sandstone and Bayfield group) of Wisconsin, Minnesota, and Michigan have cross-bedding that appears to outline the configuration of the Lake Superior syncline, is dominantly fluvial in origin, and hence closely reflects paleoslope (Hamblin, 1961, p. 6, 7, 8).

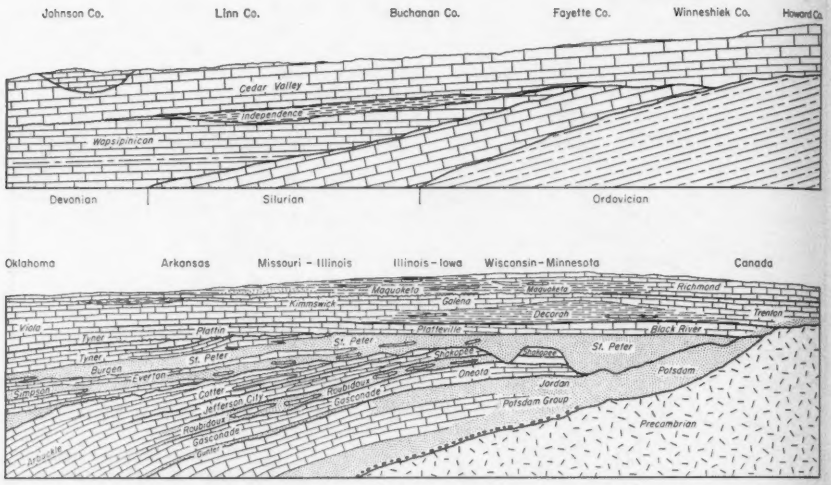


Figure 13. North-south cross sections of lower Paleozoic sediments: (A) Devonian relations in north-central Iowa (after Stainbrook, 1935, Fig. 205), (B) lower Paleozoic relations from Oklahoma to Canada (modified from Dake, 1921, Pl. 2)

The older Keweenawan cross-bedding data from upper Michigan and adjacent Wisconsin also show, along the south shore of Lake Superior, a dominant transport direction to the northwest, a direction similar to that of the overlying Cambrian sandstones of the same area. On Isle Royal on the north limb of the Lake Superior syncline cross-bedding dips to the southeast. These sandstones are believed to be largely of fluvial origin (Thwaites, 1912, p. 101–102; Hamblin and Hornor, 1961, p. 210) and closely reflect the configuration of the Lake Superior syncline.

The pre-Keweenawan quartzites in the Lake Superior region from northeastern Iowa to the northern shore of Lake Huron (Brett, 1955; Pettijohn, 1957a; McDowell, 1957) have dominantly shallow-water sedimentary structures and in some places are more than 5000 feet thick. The dominant transport direction is to the southeast. Along the north shore of Lake Huron this direction has prevailed through almost 20,000 feet of sediments (Pettijohn, 1957a, p. 477). This uniformity in current direction, both in thick section and over wide area, suggests major tectonic control. Pettijohn (1957a, p. 477) and McDowell (1957, p. 24–31) have inferred that these quartzites represent detritus derived from a positive area to the northwest.

DISPERSAL CENTERS

General

One of the most striking single characteristics of the Paleozoic, Cretaceous, and Tertiary sediments of the upper Mississippi Valley and adjacent areas is the number of relatively thin and often widely traceable sedimentary units. In Illinois alone about 170 pre-Pleistocene formations have been named. The Pennsylvanian sediments of the stable shelf area of western Illinois provide an extreme example. In this area Wanless (1957, p. 51–122) has recognized 154 members in a total Pennsyl-

TABLE 8. PROVENANCE INTERPRETATION

	Petrographic character of major source	Major dispersal center	Clastic dispersal mechanism		Sedimentary tectonics
			To basin	In basin	
Tertiary and Cretaceous	Metamorphic rocks, especially mica schists and some acid gneisses	Piedmont Plateau of ancestral southern Appalachian mountains	Single major drainage system	Delta and shallow marine shelf	Postorogenic rejuvenation and uplift of bordering tectonic lands
Pennsylvanian and Mississippian	Immature lower Paleozoic sediments and some crystalline rocks – – – – – – Mature and immature Paleozoic sediments and some crystalline rocks	Tectonic lands of the ancestral northern Appalachian Mountains	Several large southwestward-flowing drainage systems	Coastal plain and shallow marine shelf	Orogenic activity of bordering tectonic lands
Devonian	Principally Ordovician sediments		Relatively small streams draining Paleozoic hinterland	Shallow marine shelf, some littoral bordering epicontinental seas, and very minor fluvial	Stable craton
Ordovician	Principally Cambrian sediments	Lake Superior region			
Cambrian	Late and middle Precambrian sediments and crystalline rocks		Relatively small streams draining Precambrian hinterland		

vanian section of only 500 feet. Such sections contrast sharply with the thick units of orogenic sedimentation.

The presence of many widely traceable units in the upper Mississippi Valley region implies many widespread domains of sedimentation. Their thinness is the result of comparatively slow sedimentation as well as of frequent environmental change. The commonly associated cross-bedding indicates that the principal source area for the Cambrian sandstones of upper Michigan was a linear positive area extending eastward through northern Michigan and connecting the Wisconsin arch with a positive area in Ontario (Hamblin, 1958, p. 135; 1961, p. 15–18). Along the north side of this linear source area in upper Michigan, the paleoslope was initially northwest. Along the

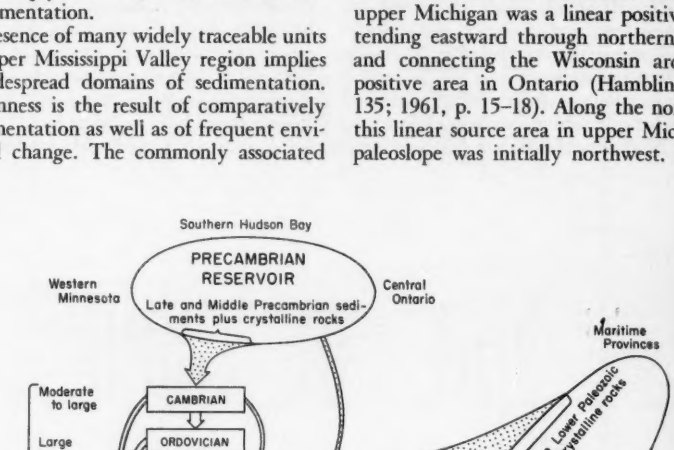


Figure 14. Dispersal centers and source relations of the post-Precambrian clastics. Stipple density is proportional to maturity of sandstones.

minor unconformities and diastems indicate many minor interruptions of sedimentation. In terms of a geologic history that emphasizes minor environmental changes and the details of regression and transgression of epicontinental seas on a stable platform, these sediments do have a complex history. But judged from the point of view of the recognition and persistence of major clastic dispersal centers, this region has a very simple history. The integrated interpretation of petrology, cross-bedding, major unconformities, and facies maps form the foundations of this unifying, internally consistent interpretation. The salient points of this interpretation are summarized in Table 7 and Figure 14.

Lower Paleozoic

The combined evidence of petrology and south side of this source region the paleoslope probably was initially oriented southward and southeastward toward the Michigan basin. Southwestward in Wisconsin the Precambrian rocks of the Wisconsin arch were an important source. Farther westward the Transcontinental arch also probably contributed clastic sediments. Excluding upper Michigan, the paleoslope was oriented dominantly to the southwest, south, and southeast throughout most of this broad region. Probably most of the sands and muds in Cambrian time were transported to shallow marine shelves by relatively small streams draining the Precambrian hinterland.

Locally abundant feldspar, garnet, and the presence of some unstable heavy minerals indicate the importance of the direct contribution of crystalline rocks to Cambrian sandstones. The relative mineralogical and textural ma-

urity of many Cambrian sandstones, however, also indicates that Keweenawan and pre-Keweenawan sandstones were important and may even have been the principal contributors.

The greater maturity of most of the Ordovician sandstones implies a source of pre-existing sediments. The pre-St. Peter unconformity, cross-bedding, and petrographic data show that most of the Ordovician sandstones were derived from the underlying Cambrian sandstones. Although studies of heavy minerals show some unstable species, especially where the Ordovician is proximal to the Precambrian basement, direct crystalline rock contributions to Ordovician sandstones appear to have been minor. Relatively small streams draining the Cambrian hinterland probably deposited most of the sands and muds on shallow, stable marine shelves. The pre-St. Peter unconformity and the available cross-bedding indicate that the paleoslope was dominantly to the south, southeast, and southwest throughout much of the central part of the craton.

The extreme mineralogical and textural maturity of the Devonian sandstones implies derivation from mature pre-existing sediments. The Ordovician sandstones were the principal source and the Cambrian sandstones a probable minor source. Crystalline rock contributions were almost nonexistent.

The petrologic properties of the lower Paleozoic sandstones are therefore the result of progressive sedimentary differentiation in a sedimentation system to which crystalline rock contributions were only initially important. Successive rejuvenation of the craton to the north was responsible for persistent southerly paleoslopes and led to the progressive recycling of Cambrian, Ordovician, and Devonian sandstones under conditions of extreme tectonic stability. Be they Devonian, Ordovician, or Cambrian in age, the pre-Mississippian sandstones of the upper Mississippi Valley and adjacent areas all have the same ultimate source, the reservoir of Precambrian rocks centered about the Lake Superior region.

Mississippian and Pennsylvanian

Beginning with Mississippian time, the importance of the Lake Superior dispersal center diminished. For the first time in the upper Mississippi Valley and adjacent areas, major amounts of immature clastics were derived from east of the Cincinnati and Findlay arches. The key indices of this new source region are polycrystalline quartz, angular grains, and abraded tourmaline overgrowths on abraded cores.

Available subsurface and cross-bedding data indicate that most immature Osage and Kinderhook clastics of the Illinois and Michigan basins were derived from the northeast. Petrographically similar sandstones of broadly comparable age (Waverly, Cuyahoga, Berea, and Pocono) occur in the Appalachian basin in Ohio, Kentucky, West Virginia, and Pennsylvania. Pepper *et al.* (1954, p. 95–107, Fig. 59) have indicated that the dominant source of the Berea sandstone in northern Ohio lay to the northeast and hypothesized that an ancient river system drained southwestward and collected clastics from both the Canadian Shield and the northern Appalachian basin. In western Pennsylvania and West Virginia they also recognized contributions from the east and southeast. In Pennsylvania the Pocono deltaic sequence was derived from tectonic borderlands southeast of the present Appalachian basin (Pelletier, 1958, p. 1056–1059). The large volume of these immature sandstones and the known middle and late Paleozoic activity of the tectonic borderland southeast of the ancestral northern Appalachian area minimize the role of Canadian Shield contributions. Another factor favoring a predominant Appalachian source is the abraded tourmaline overgrowths on abraded cores. If these had originated in the Precambrian rocks of the Canadian Shield they probably would not be restricted to Mississippi and younger sands. Although we recognize the role of Canadian Shield contributions, we believe that a large southwestward-flowing river system, collecting sands and muds mainly from the northern Appalachian area, was the principal source of the Osage and Kinderhook subgraywackes of the Illinois and Michigan basins. Comprehensive petrographic studies of the sandstones coupled with age determinations of detrital feldspars and micas such as have been made by Vistelius (1959) would help evaluate the importance of Canadian Shield contributions.

A similar river system flowing southwest but shifting laterally as time passed is the most plausible means of distributing the Chester clastics of the Illinois basin. The maturity of the Chester sandstones suggests some contributions from the lower Paleozoic sandstones of the upper Mississippi Valley region.

The Pennsylvanian sandstones also initially received important contributions from the lower Paleozoic and Mississippian sandstones of

the upper Mississippi Valley and adjacent regions. Increasing immaturity, large volume, and the available studies of cross-bedding and pebble gradient (Potter and Siever, 1956, p. 240–243; Nickelsen, 1958, p. 791–796; Pelletier, 1958, p. 1059–1061) show, however, that highlands southeast of the northern half of the Appalachian basin contributed most of the clastics, although at least some were transported southwestward parallel to the axis of the geosyncline (Williams, 1959, Fig. 4). These clastics entered the upper Mississippi Valley region from the east and northeast. There is no petrographic evidence to support the idea that the younger Pennsylvanian subgraywacke sandstones had as an important source the older Pennsylvanian protoquartzites of the now eroded western flank of the Cincinnati arch (Friedman, 1960, p. 40–45). Smaller streams entering the Illinois and Michigan basins from the north and northwest also made contributions.

Thus the great bulk of the Mississippian and Pennsylvanian clastic sediments of the Illinois and Michigan basins reflects contributions from a dispersal center located along the southeastern flanks of the northern half of the Appalachian geosynclinal belt. Pre-existing immature sediments and some crystalline rocks were the chief source rocks. Several large drainage systems heading in the northern Appalachian region supplied sands and muds to a combined broad coastal plain and shallow marine shelf in the Illinois and Michigan basins.

Cretaceous and Tertiary

During the erosional interval between Pennsylvanian and Cretaceous time in the upper Mississippi Valley region the principal dispersal center shifted southward.

The abundance of metamorphic heavy minerals such as kyanite, sillimanite, and staurolite in the Cretaceous and Tertiary sands of the upper portions of the Mississippi Embayment indicates that most of the sediments were derived from the Piedmont regions of the ancestral southern Appalachian mountains (Pryor, 1960). Although some grains of kyanite, sillimanite, and staurolite have been noted in some of the Paleozoic sandstones of the upper Mississippi Valley, the abundance of these minerals in Cretaceous and Tertiary sands shows that the mica schists and gneisses of the Blue Ridge and Piedmont Plateau region were the major source. This conclusion, in conjunction with evidence of facies distribution and paleoslope, indicates that the principal dispersal center of the Cretaceous-Tertiary clastics of the upper parts of the Mississippi Embayment was the rejuvenated and deeply eroded axial parts of the southern part of the Appalachian geosyncline. Some Paleozoic sedimentary rocks, probably mainly Pennsylvanian, also made small contributions. A major drainage system heading in the Piedmont region transported most of these sands and muds to a combination delta and shallow marine shelf at the northern end of the Mississippi Embayment. Essentially the same source and paleoslope relations prevailed in the Tertiary. Cretaceous and nearby pre-Cretaceous sands and clays also made some minor contributions to the Tertiary sediments.

To summarize, the Paleozoic, Cretaceous, and Tertiary clastics of the upper Mississippi Valley and adjacent areas have had three major dispersal centers: the Precambrian rocks of the Lake Superior region, highlands along the southeast side of the ancestral northern Appalachian mountains, and highlands along the southeast side of the ancestral southern Appalachian mountains. In broadest terms, the pre-Mississippian sandstones of this part of the craton were all ultimately derived from the Lake Superior dispersal center. In contrast, most of the Mississippian and later clastics are related to Appalachian orogenic activity.

A series of large, recurring drainage systems, whose positions and orientations appear to have been controlled chiefly by regional tectonic elements, transported most of the sands and muds from the Appalachian dispersal centers to the low-lying coastal plains, shallow marine shelves, and small deltas of the post-Devonian sedimentation of the craton.

PROBLEMS OF REGIONAL AND METHODOLOGICAL INTEREST

General

Inherent in the foregoing results and interpretations are a number of problems, some of regional and others of more general interest. Our study has contributed to the solution of some of these. It has also brought into sharp focus those whose precise solution is yet to be found. In this section we comment briefly on these aspects of our study.

Sedimentary Differentiation

Sedimentary differentiation refers to the systematic changes in the mineralogy and texture of sandstones that result from pro-

gressive erosion, abrasion, and sedimentation. These changes—progressive sorting, rounding, and elimination of unstable detrital minerals—make it possible in theory to derive each major sandstone type from crystalline rock. Study of ancient sandstones, modern sands, and laboratory experiments have contributed to this concept. Although not commonly stated, assumptions concerning the rate or magnitude of sedimentary differentiation play on important role in the interpretation of source areas and of the environments of sandstones. Because of both its general importance and its particular bearing on the source relations of the sandstones of the upper Mississippi Valley and adjacent areas, we point out the underlying assumption of this paper, present an alternative, and indicate a possible future solution.

Our evaluation of sedimentary differentiation is indicated by our interpretation of the progressive increase in maturity of the lower Paleozoic sandstones. We consider this the result of progressive sedimentary differentiation in a sedimentation system to which crystalline rocks were only initially important in Cambrian time. Inherent in this interpretation is the assumption that mature and supermature sandstones cannot be directly produced in large volumes from crystalline rocks under the usual conditions of erosion and sedimentation that prevailed in the geologic past. Studies of modern sands and experimental work support this assumption. Fluvial transport rounds sand grains very slowly (Kuenen, 1959). Although differential abrasion can be produced experimentally, major mechanical wear never has been demonstrated in large natural streams (Russell, 1939, p. 44–45; Van Andel, 1950, p. 114; Pollack, 1961, p. 35–36). Modern beaches and dunes, although containing better rounded sands than their local sources (Beal and Shepard, 1956, Fig. 2), also fail to produce sands either as well rounded or as mineralogically stable as those of the St. Peter and Devonian sandstones.

The alternate assumption concerning sedimentary differentiation is that under conditions of prolonged tectonic quiescence crystalline rocks can directly yield mature and supermature sands, so that little or no trace is left of intermediate, less mature abrasion products. Such an assumption implies that in the geologic past extremely different conditions at least occasionally prevailed that have no duplicates today. For example, Dake (1921, p. 217–218) suggested that the sand grains of the well-rounded lower Paleozoic sandstones may have been largely the result of eolian processes acting in a sparsely vegetated source region. According to this assumption the Ordovician and Devonian sandstones of the upper Mississippi Valley could have been derived directly from the crystalline rocks of the Lake Superior dispersal center rather than chiefly from the progressive recycling of pre-existing sandstones.

Folk (1960, Fig. 3, p. 22, 24, 35) is a modern advocate of the effectiveness of sand rounding in environments with high kinetic energy. In the Silurian sandstone of West Virginia he found evidence to support the view that appreciable rounding can be obtained in a single cycle. His interpretation is based on the qualitative roundness contrasts of intimately interbedded sandstone units—the well-rounded sand grains are inferred to be of beach origin and the less rounded ones of estuarine origin. Although his interpretation may indeed be correct, no independent evidence from heavy minerals, sorting, or sedimentary structures is offered to support it.

In the upper Mississippi Valley region, evidence for direct derivation of the Ordovician and Devonian sandstones from crystalline rocks is not compelling. There the known availability of Cambrian sandstones and the pre-St. Peter unconformity minimize the need for invoking such a hypothesis. In addition, the extreme maturity of the heavy-mineral suite of the Ordovician and Devonian sandstones strongly argues against a first-cycle origin. Instead, available evidence indicates that the mature Ordovician sandstones were derived principally from the Cambrian sandstones, and the supermature Devonian sandstones were derived principally from a major Ordovician and a minor Cambrian source. Nor need most of the Cambrian sandstones have been directly derived from Precambrian crystalline rocks. The availability of Precambrian sedimentary rocks in the Lake Superior region and the relative maturity of the heavy-mineral suite in most Cambrian sandstones minimize the necessity of inferring a first-cycle origin for the bulk of the Cambrian sandstones.

Although we favor the "recycling hypotheses" for the sandstones of this study, we do not believe that the Paleozoic and later sandstones of the upper Mississippi Valley can provide a general solution to this problem. Because of the uncertainty introduced by presumed variations in time and intensity factors, studies of modern sands and experimental work probably will

contribute only supporting evidence. Studies of ancient mature sandstones in other basins and areas do hold promise of an answer. The correlation between polycrystalline quartz and grain angularity shown in Figure 6 illustrates this point of view.

The conventional interpretation of Figure 6 is that the greater grain angularity and abundance of polycrystalline quartz in the post-Devonian sandstones reflect immature sands derived from a source region that had been subjected to more intense local and regional deformation than had the pre-Devonian source. An alternate interpretation of Figure 6 is that, under the conditions of abrasion required to produce mature and supermature sandstones, polycrystalline quartz disintegrates along grain boundaries, moves into smaller size classes, and becomes indistinguishable from other smaller nonpolycrystalline quartz grains. The durability of polycrystalline quartz is the basic issue underlying these alternatives. Is polycrystalline quartz as durable as monocrystalline quartz or will it disintegrate under conditions of long and/or intense abrasion? Although definitive evidence is lacking, Folk (1960, p. 24) suggests that polycrystalline quartz has nearly the same abrasion resistance as common quartz. Abundance of polycrystalline quartz in other sandstones with well-rounded quartz grains would easily resolve this problem.

The systematic examination of other basins or regions that contain mature and supermature sandstone would thus seem to offer the most direct and promising means of evaluating the sedimentary differentiation.

Mississippi Embayment and Illinois Basin

An impressive parallel between the Tertiary and Cretaceous sediments of the upper Mississippi Embayment and those of Chester and Pennsylvanian time in the Illinois basin is apparent from our studies. Particularly striking are the similarities between paleoslope, transport direction and depositional strike, lithologies, spatial distribution of marine and nonmarine sediments, unconformities, and broad physiographic settings of the two groups.

The Tertiary and Cretaceous sediments are separated by a major unconformity that truncates older sediments to the north, northeast, and northwest. The Pennsylvanian and Chester sediments are similarly separated. The Eocene sediments of the Tertiary contain fewer marine units than the Cretaceous and consist dominantly of nonmarine sands and muds with subordinate lignitic beds. A like contrast exists between Pennsylvanian and Chester. Nonmarine clastics and coal beds are many times more prevalent in the Pennsylvanian than in the Chester.

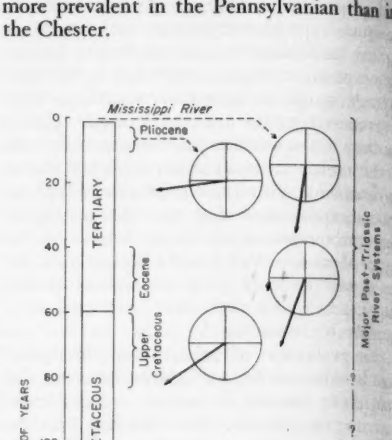


Figure 15. Transport directions and time scale of the post-Devonian sandstones of the Illinois basin and upper Mississippi Embayment

Most of the clastics of the four periods were supplied by large integrated drainage systems whose positions and orientations were largely tectonically localized. For all four periods regional transport direction was dominantly down the paleoslope at right angles to depositional strike. Both basins were filled dominantly by clastics entering at the northern end. Because of this, marine sediments were similarly distributed within the basins and became thicker downdip. Together these facts indicate that the Tertiary and Cretaceous sediments of the upper Mississippi Embayment and the

Chester and Pennsylvanian sediments of the Illinois basin belong to the same sedimentary association and have the same general mode of deposition. Pryor (1960, Table VIII) outlined a model for the late Cretaceous sediments of the Mississippi Embayment that, with some modification, also applies to the above Eocene, Pennsylvanian, and Chester sediments.

Three points in this comparison deserve emphasis.

First, from at least early Mississippian time the area between the Cincinnati arch and the Ozark dome–Mississippi arch has served as a sluiceway (Fig. 15) in which clastic materials were transported toward the continental margin. Although broad cross-folds such as the Pascola and Kankakee arches developed, they appear to have little or no effect on clastic transport pattern. Throughout portions of Mississippian and Pennsylvanian time this sluiceway was undoubtedly an important sediment contributor to a miogeosynclinal belt connecting the Appalachian and Ouachita basins.

Second, there was a dominance of longitudinal filling (Fig. 15) in this elongate, weakly negative area because of the tectonic localization of several ancient river systems. Our study thus confirms the deductive conclusion of Kuenen (1957, p. 194), who has stressed the longitudinal filling of many oblong basins and embayments.

Third, because of similarities in tectonic setting, basin geometry, and mechanism of clastic filling, the sediments of the upper part of the Mississippi Embayment are a Cretaceous-Tertiary analogue, differing only in scale, of the Chester-Pennsylvanian sediments of the Illinois basin. The two basins differ chiefly in their relation to the present continental margin. The upper Mississippi Embayment is connected with an integral part of the present continental margin, whereas the Illinois basin has long since been separated from it. Thus the upper Mississippi Embayment can be thought of as a "modern" intracratonic analogue of the Illinois basin in Mississippian and Pennsylvanian time.

Intrastratal Solution

We have explained the increasing complexity of heavy-mineral suites in younger sediments with the mechanism of changing source areas or dispersal centers. However, an alternative possibility for this increased complexity does exist and should be considered. This alternative mechanism is the postdepositional destruction of minerals, or intrastratal solution (Pettijohn, 1941). In this concept the older sediments would have progressively lost the less stable mineral species by selective removal in the deeper zones through intrastratal solution. Krynine (1942) and van Andel (1959) have taken exception to the wide application of this concept, and their objections to intrastratal solution are applicable in the case of the Paleozoic and later clastics of the upper Mississippi Valley and adjacent areas.

Cambrian, Ordovician, and Mississippian sediments locally contain, in addition to the ubiquitous zircon and tourmaline, minor quantities of less stable mineral species, such as hornblende, epidote, kyanite, staurolite, monazite, topaz, sphene, actinolite, and tremolite. These minerals do not exhibit corrosive structures that should be expected of remanent grains of a mineral suite attacked by intrastratal solution.

On the basis of the presence of some less stable heavy mineral species in Cambrian, Ordovician, and Mississippian sediments and the arguments advanced against intrastratal solution by Krynine (1942) and especially by van Andel (1959, p. 156–160), we conclude that intrastratal solution has not played a major role in the variation of heavy-mineral suites in the sediments of this area.

Clastic Sedimentation and the Craton

In the approximately 500 million years since the beginning of the Cambrian, marine transgressions and regressions of the craton have been numerous. They have left behind a record of many thin units, traceable for long distances, and many minor and a few major unconformities. The sandstones of this cratonic sequence all reflect deposition in dominantly shallow water.

Instead of the bewildering array of different transport directions that might be expected from sands and muds in shallow epicontinental seas washed more or less randomly across the craton, the available evidence indicates a strong directional unity in all these sandstones, be they of dominantly marine-shelf or continental origin. In the sample area the dominant transport direction has been the same for nearly 500 million years. Such uniformity over so long a time and over such a wide area can reflect only major tectonic control. The behavior of basement rocks of the craton provides that control. This underlying tectonic control is the

immediate cause of persistent paleoslopes, of recycling, and of the localization and orientation of major clastic deposits ultimately derived from distant tectonic lands.

The relationships between the pattern of truncation along major unconformities and transport directions in the underlying and overlying sandstones are noteworthy. Along the south shore of Lake Superior the cross-bedding of both the Keweenawan and overlying Cambrian sandstones is oriented northwestward. Along this major unconformity the Jacobsville sandstone thins in the updip cross-bedding direction (Hamblin, 1958, Fig. 2). In the Cambro-Ordovician, the pre-St. Peter unconformity, with its truncation of older sedimentary rocks to the north, has southwestward-oriented Franconia and Trempealeau cross-bedding below it and southwestward-oriented St. Peter cross-bedding above it. Similar relationships are well documented in the Chester and basal Pennsylvanian sandstones that have nearly identical average cross-bedding orientations (215° and 214°) and are separated by a major unconformity that truncates older rocks to the northwest, north, and northeast across the basin. The unconformity separating Cretaceous from Tertiary in the upper Mississippi Embayment bears a similar relation to the orientation of overlying and underlying cross-bedding. Cretaceous and Eocene cross-bedding is oriented toward the southwest, older Cretaceous sediments are truncated toward the embayment margins, and the Eocene forms a series of southward offlapping units. Although marine-shelf sands are subordinate in the Tertiary, Cretaceous, Pennsylvanian, and upper Mississippian, the sandstones above and below the pre-St. Peter unconformity are dominantly marine-shelf sands.

The above evidence shows that paleoslopes may control sand dispersal, the pattern of truncation of associated unconformities, and direction of onlap and offlap relations as well. From knowledge of any one of the three it may be possible to infer the other two. Collectively, these relations form the essential key to the growth patterns of the sediments veneering the craton throughout most of its post-Precambrian history.

Figure 16 shows the essential steps of overlap and regression, followed by renewed overlap or overstep. This sequence is the result of repeated rejuvenation of the older, more stable portion of the craton. Unconformities become compound the more closely one approaches the central stable region. Repetition of this pattern throughout long periods of time on a predominantly southwestward paleoslope is the basic explanation for cross-bedding orientation, the observed pattern of truncation at major unconformities, most of the major facies changes observed on the craton, and recycling.

In the absence of important contributions from a marginal tectonic belt, clastic materials are derived from the craton itself and recycled under conditions of extreme stability. When sands and muds from a tectonic borderland

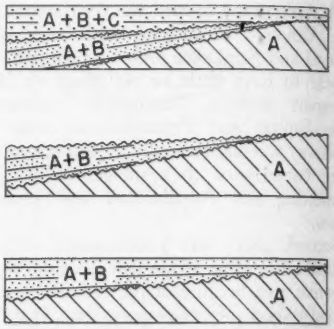


Figure 16. Diagrammatic representation of overlap and overstep, unconformities, and resultant recycling

reach the craton, their dispersal pattern is largely controlled by the slope of the craton rather than the tectonic strike of the bordering mobile belt. Recycling continues to prevail, but chiefly with sands and muds originally derived from tectonic borderlands. The development of intracratonic basins provides only relatively local sediment traps for clastic materials ultimately in transit to the continental margins.

SUMMARY AND CONCLUSIONS

Dispersal centers of the Paleozoic and later clastics of the upper Mississippi Valley and adjacent areas were investigated. Sedimentary petrology and directional evidence based on cross-bedding, facies distribution, and patterns of truncation shown by major unconformities were used to locate the three principal dispersal centers of this area of the craton.

Specific petrologic conclusions include the following:

(1) Angularity of tourmaline grains cor-

relates closely with that of other principal detrital components.

(2) Tourmaline angularity correlates with abundance of polycrystalline quartz.

(3) The predominant heavy minerals of the Paleozoic sandstones are zircon and tourmaline with garnet as an important accessory, whereas the predominant heavy minerals of the Cretaceous and Tertiary sands are kyanite, staurolite, and sillimanite.

(4) Illite is the chief component in Paleozoic clay-mineral assemblages, whereas montmorillonite and kaolinite are the chief components of the Cretaceous and Tertiary sediments.

(5) Devonian and Ordovician sandstones are predominantly orthoquartzites, whereas feldspathic sandstones are present in the Cambrian.

(6) The great bulk of Mississippian and Pennsylvanian sandstones are protoquartzites and subgraywackes.

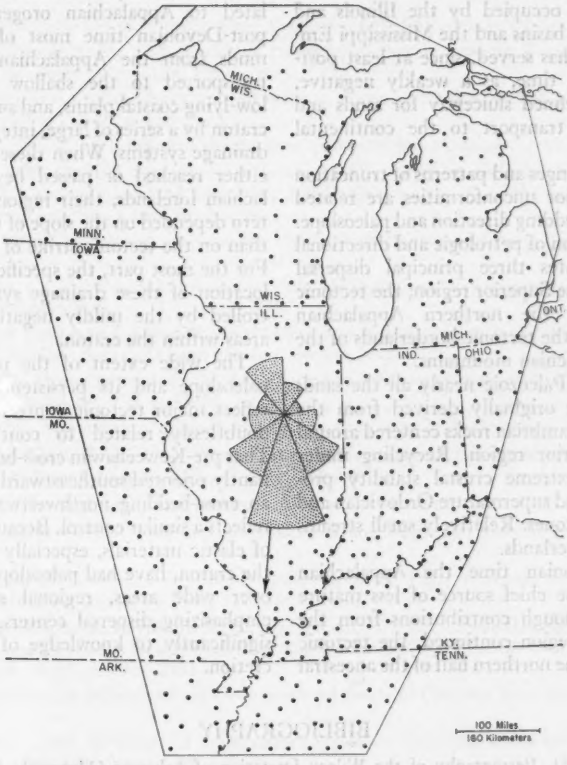


Figure 17. Sample points, windrose, and grand mean of tangents to present streams

(7) The Cretaceous and Tertiary sandstones are protoquartzites.

Specific conclusions concerning directional evidence include the following:

(1) Pre-Keweenawan (middle Precambrian) cross-bedding is generally oriented to the southeast.

(2) Keweenawan (late Precambrian) cross-bedding is generally oriented to the northwest.

(3) Paleozoic and later cross-bedding is generally oriented to the southwest.

(4) The average value of all 9141 pre-

Pleistocene cross-bedding measurements is 194° and differs only slightly from the average direction, 188°, of the region's stream pattern.

(5) Most of the major unconformities of the region truncate older sediments to the north, northwest, and northeast.

(6) The area occupied by the Illinois and Michigan basins and the Mississippi Embayment has served, since at least post-Devonian time, as a weakly negative, clearly defined sluiceway for sands and muds in transport to the continental margin.

(7) Facies changes and patterns of truncation along major unconformities are related to cross-bedding direction and paleoslope.

The integration of petrologic and directional evidence indicates three principal dispersal centers: the Lake Superior region, the tectonic borderlands of the northern Appalachian mountains, and the tectonic borderlands of the southern Appalachian mountains.

In the lower Paleozoic nearly all the sands and muds were originally derived from the reservoir of Precambrian rocks centered around the Lake Superior region. Recycling under conditions of extreme crustal stability produced mature and supermature Ordovician and Devonian sandstones. Relatively small streams drained the hinterlands.

In post-Devonian time the Appalachian reservoir was the chief source of less mature sandstones. Although contributions from the Lake Superior region continued, the tectonic borderlands of the northern half of the ancestral Appalachian mountains supplied most of the Mississippian and Pennsylvanian clastic materials. In the Cretaceous and Tertiary, the ancestral southern Appalachian region, now deeply eroded and tectonically inactive, was the dominant source.

Most of the post-Devonian clastics are related to Appalachian orogenic activity. In post-Devonian time most of the sands and muds from the Appalachian reservoir were transported to the shallow marine shelves, low-lying coastal plains, and small deltas on the craton by a series of large, integrated, recurring drainage systems. When these sands and muds either reached or passed beyond the Appalachian forelands, their regional dispersal pattern depended on the slope of the craton rather than on the tectonic strike of the mobile belt. For the most part, the specific orientation and location of these drainage systems were controlled by the mildly negative and positive areas within the craton.

The wide extent of the post-Precambrian paleoslope and its persistence through time reflect major tectonic control. This control is doubtlessly related to continental growth. That pre-Keweenawan cross-bedding was dominantly oriented southeastward and Keweenawan cross-bedding northwestward undoubtedly reflects a similar control. Because large volumes of clastic materials, especially those veneering the craton, have had paleoslopes that persisted over wide areas, regional sediment study, emphasizing dispersal centers, can contribute significantly to knowledge of continental accretion.

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Appendix 1. Tourmaline Angularity

Because we wished to determine the degree to which the angularity of tourmaline grains correlated with the angularity of the other principal detrital elements, we have used the multiple correlation coefficient. The multiple correlation coefficient expresses the correlation between one variable, for example ξ_1, and the totality of the other variables, for example $\xi_2\ \xi_3\ \ldots\ \xi_n$, or any subset of them, $\xi_2\ \xi_3 \ldots \xi_k$ ($k<n$), in the system. Expository accounts of the multiple correlation coefficient are given in many elementary statistical textbooks, and Cramer (1946, p. 301–331), Anderson (1958, p. 86–96), and others derive theory and tests of significance. Although the multiple correlation coefficient is widely used in other fields, comparatively little use has been made of it in geology. Jacobsen (1959, p. 77) and Hubert (1960, p. 208–211), however, give examples of its use in petrographic studies of sandstones.

The data of our experiment were obtained by randomly selecting two sandstone samples from each of the seven systems included in the study and determining the percentage of angular tourmaline, zircon, and quartz grains of the .125- to .177-mm and .062- to .500-mm classes. Percentages of angular tourmalines and zircons of the .062- to .500-mm fraction were estimated from heavy-mineral mounts and the percentage of angular quartz from loose grains mounted on glass slides. Estimates were based on counts of 100 grains. The identity of the slides was unknown to the operator. The tabulation shows the resultant data.

Letting ξ_1 = percentage of angular tourmaline grains (.062–.500 mm)
ξ_2 = percentage of angular zircon grains (.062–.500 mm)
ξ_3 = percentage of angular quartz grains (.125–.177 mm)
ξ_4 = percentage of angular quartz grains (.062–.500 mm)

and using the arc sine transformation for percentages, the data were programmed for the Illiac digital computer yielding a correlation matrix and multiple and partial correlation coefficients.

The resultant matrix is

$$\begin{bmatrix} 1.000 & 0.845 & 0.800 & 0.877 \\ 0.845 & 1.000 & 0.794 & 0.916 \\ 0.800 & 0.794 & 1.000 & 0.916 \\ 0.877 & 0.916 & 0.916 & 1.000 \end{bmatrix}$$

which summarizes the correlation observed in the sample between ξ_1 and ξ_2, ξ_1 and ξ_3, and so on. Thus 1.000 is the correlation between per cent of angular tourmaline and itself, 0.845 is the correlation between angularity of tourmaline and zircon, 0.794 between angularity of zircon and .125 to .177 mm quartz, and so on.

The multiple correlation coefficient between ξ_1 and ξ_2, ξ_3 and ξ_4 is then given by

$$\hat{\rho}_{1\cdot 234} = \sqrt{1 - \frac{\hat{P}}{\hat{P}_{11}}}$$

where $\hat{P}$ is the sample correlation matrix and $\hat{P}_{11}$ is the appropriate partition.

The multiple correlation between tourmaline angularity and that of zircon, .125–.177 mm quartz and .062–.500 mm quartz was 0.8834. The multiple correlation between tourmaline, zircon, and .062–.500 mm quartz angularity was 0.8831, almost identical.

To test the hypothesis that tourmaline angularity is independent of the angularity of the other three components we compute (Anderson, 1958, p. 93)

$$\frac{\hat{\rho}^2_{1\cdot 234}}{1-\hat{\rho}^2_{1\cdot 234}} \left(\frac{N-p}{p-1}\right)$$

where N is the number of observations and p the number of variables. If this quantity is greater than $F_{p-1,N-1}(0.01)$, we reject the hypothesis. As the computed value is 13.39 and exceeds the critical value of $F_{3,10}(0.01) = 6.55$, we reject the hypothesis that tourmaline angularity is independent of the angularity of the other principal detrital elements in the sandstones of this study. This experiment demonstrates the degree to which tourmaline angularity correlates with that of the other principal detrital components of the sandstones of this study. Doubtlessly, similar correlations prevail in most other sandstones.

Because most properties of sedimentary rocks represent the intercorrelated response to the broad controls of provenance, depositional environment, and diagenesis, the multiple correlation coefficient can be used to solve many problems. For example, one might be interested in the multiple correlation between bed thickness and a collection of other related variables such as grain size, sorting, and presence or absence of cross-bedding. Or one might be interested in the multiple correlation between the abundance of a trace element and that of the other trace elements in a given sedimentary environment. Moreover, any variable of a set can be correlated with the others or any subset of the others without specifying which variables are independent or dependent. Finally, once either the correlation or covariance matrix of a set of data is obtained, it is relatively simple to compute the desired partial correlation coefficients or to obtain regression estimates of the dependence of one variable on the others in the system.

PERCENTAGE OF ANGULAR GRAINS IN SAMPLES OF SYSTEMS STUDIED

Samples	Tourmaline (.062–.500 mm)	Zircon (.062–.500 mm)	Quartz (.125–.177 mm)	Quartz (.062–.500 mm)
Tertiary	96	79	99	95
Tertiary	82	69	95	95
Cretaceous	90*	85	100	97
Cretaceous	37	55	87	74
Pennsylvanian	100	71	89	98
Pennsylvanian	83	59	86	87
Mississippian	40	63	84	84
Mississippian	66	78	60	91
Devonian	06†	08	09	01
Devonian	05	07	12	14
Ordovician	04	27	02	04
Ordovician	01	26	56	48
Cambrian	41	53	28	58
Cambrian	41	24	75	56

*47 grains
†68 grains

APPENDIX 2. PETROGRAPHIC ANALYSES

PETROGRAPHIC ANALYSES OF SAMPLES FROM THE SYSTEMS STUDIED
(Per cent)

Sample	Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Rock fragments	Miscellaneous	Polycrystalline quartz	Tourmaline Round	Tourmaline Angular
					Tertiary						
680	92.0	3.0	0.0	0.0	2.0	3.0	0.0	0.0	9.5	2	11
681	82.0	13.5	0.0	0.5	2.5	1.0	0.0	0.5	9.0	..	..
682	77.0	9.0	0.0	1.5	2.5	7.5	0.5	1.0	10.5	..	..
683	86.0	9.0	0.0	1.0	3.0	1.0	0.0	0.0	13.5	..	..
684	83.5	4.0	0.0	1.0	1.0	2.0	0.0	0.0	9.5	..	..
685	73.0	16.0	0.0	0.0	1.0	8.0	2.0	0.0	11.5	..	..
					Upper Mississippian (Chester)						
9*†	55.0	29.0	11.0	0.5	3.5	0.0	0.0	1.0	0.0	48	52
30*†	81.5	10.5	5.0	0.5	0.0	0.5	2.0	0.0	5.5	..	..
31*†	91.0	4.0	0.5	1.0	0.0	1.0	2.5	0.0	3.0	..	..
32*†	73.0	0.0	26.5	0.0	0.0	0.0	0.0	0.5	3.0	39	61
33*†	72.0	24.5	0.5	2.5	0.0	0.0	0.0	0.5	4.5	..	..
34*†	85.5	5.0	5.5	0.0	0.5	0.0	3.5	0.0	2.0	27	73
35*†	86.5	5.5	2.5	0.5	0.5	0.0	3.5	1.0	1.0	..	..
36*†	68.0	10.5	13.0	0.5	1.0	0.0	5.0	2.0	4.0	..	..
51†	87.5	6.0	0.5	0.5	0.5	0.0	4.0	1.0	1.0	17	83
52†	81.0	6.0	0.5	1.0	2.0	0.0	5.0	4.5	8.0	..	..
53†	91.5	6.5	0.0	0.0	0.5	0.0	0.5	1.0	3.0	..	..
55†	92.5	4.5	0.0	0.0	0.0	0.0	0.5	2.5	2.5	..	..
56†	81.5	6.5	1.0	0.5	0.0	0.0	3.5	7.0	2.0	..	..
368	94.5	4.0	0.0	0.0	0.0	0.0	0.5	1.0	3.0	50	50
370	72.0	18.0	0.0	0.0	0.0	0.0	2.0	8.0	2.0	37	63
371	76.5	16.0	0.5	0.0	0.5	0.0	4.5	2.0	2.5	39	61
372	97.0	2.5	0.5	0.0	0.0	0.0	0.0	0.0	2.5	43	57
373	74.0	4.0	19.5	0.0	0.0	0.5	1.5	0.5	0.5	29	71
374	96.0	2.0	0.5	0.0	0.0	0.0	1.5	0.0	1.0	63	37
426*	78.5	7.0	12.5	0.0	0.0	0.0	1.0	1.0	1.5	48	52
427*	89.5	6.5	0.5	1.0	0.0	0.5	2.0	0.0	3.0	34	66
429*	87.5	9.0	1.5	1.5	0.0	0.0	0.5	0.0	2.0	32	68
448	95.5	3.5	0.0	0.0	0.0	0.5	0.0	0.5	2.5	60	40
465	77.5	19.0	0.0	0.5	0.0	0.5	1.0	1.5	3.0	..	..

APPENDIX 2. *Continued*

Sample	Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Rock fragments	Miscellaneous	Polycrystalline quartz	Tourmaline Round	Tourmaline Angular
473	88.5	8.5	0.5	1.0	0.0	0.5	0.5	0.5	..	32	68
474*	83.5	4.0	12.5	0.0	0.0	0.0	0.0	0.0	5.0	39	61
475*	92.0	2.0	3.0	1.0	0.0	0.0	2.0	0.0	2.0	52	48
476	73.5	22.5	3.5	0.0	0.0	0.0	0.0	0.5	0.5	52	48
479	85.5	8.5	0.0	1.0	0.0	0.5	2.5	2.0	3.0	26	74
480	55.5	0.0	43.0	0.5	0.0	0.5	0.0	0.5	0.0	44	56
481	98.5	0.5	0.0	0.5	0.0	0.5	0.0	0.0	5.5	48	52
482	73.5	22.0	0.0	1.5	0.5	0.5	2.0	0.0	5.5	44	56
483	93.0	6.0	0.0	1.0	0.0	0.0	0.0	0.0	3.0	39	61
484	88.5	7.5	0.5	0.0	0.0	0.0	1.5	2.0	2.0	34	66
485	65.5	5.0	29.5	0.0	0.0	0.0	0.0	0.0	3.0	..	..
487	98.0	0.5	0.0	0.0	0.0	0.5	0.0	1.0	3.5	..	..
573	81.5	16.0	0.0	0.0	0.5	0.0	1.5	0.0	3.5	9	91
574	88.0	9.5	0.0	1.5	0.0	0.0	1.0	0.0	2.0	40	60
575	94.5	3.5	0.0	0.5	0.0	0.0	1.0	0.5	1.0	60	40
577	88.0	6.5	0.0	1.5	0.5	0.0	1.0	2.5	1.5	46	54
581*	90.0	4.5	5.5	0.0	0.0	0.0	0.0	0.0	1.0	..	..
582*	85.5	8.0	1.5	3.5	0.0	0.5	1.0	0.0	1.0	..	..
583*	85.5	6.5	3.0	1.5	0.5	0.0	2.5	0.5	3.5	..	..
584*	7.0	19.5	2.5	1.0	0.0	0.0	0.5	0.5	2.5	..	..
585*	53.0	1.0	44.5	0.0	0.0	0.0	0.5	1.0	4.0	..	..
586*	51.5	0.5	48.0	0.0	0.0	0.0	0.0	0.0	4.5	..	..
587*	90.5	3.0	3.5	0.5	0.0	0.5	1.5	0.5	2.5	..	..
588*	86.5	9.5	1.0	1.0	0.5	0.0	1.5	0.0	2.0	..	..
4-284*†	82.0	2.5	6.0	3.0	0.0	0.0	6.0	0.5	6.0	..	..
4-300*†	95.0	1.0	2.0	0.0	0.0	0.5	1.0	0.5	0.5	..	..
4-326*†	97.0	0.5	2.0	0.0	0.0	0.5	0.0	0.0	1.5	40	60
4-334*†	70.0	29.0	0.0	0.0	0.0	0.0	0.0	1.0	6.5	..	..
C-1319*†	94.0	2.5	2.5	0.5	0.0	0.0	0.0	0.5	3.5	36	64
C-1488*†	94.0	2.0	0.0	0.5	0.0	0.0	0.5	3.0	4.5	50	50
C-1803*†	91.5	2.0	0.5	0.0	0.0	0.0	2.0	4.0	2.5	33	67
C-1974*†	93.0	3.5	0.5	2.0	0.0	0.0	1.0	0.0	2.0	..	..
C-2179*†	92.5	4.5	0.5	0.5	0.0	0.0	2.0	0.0	4.0	47	53
				Mississippian (Middle and Lower)							
369*	88.5	1.0	4.5	0.0	0.0	3.5	0.0	2.5	1.5	..	..
420	88.0	10.0	0.5	0.0	0.0	0.0	0.5	1.0	0.5	36	64
424	53.5	0.0	46.5	0.0	0.0	0.0	0.0	0.0	0.0	103	6
428*	71.5	11.5	2.5	1.0	8.0	0.0	2.5	3.0	1.0	1	83
432	88.0	3.0	6.0	0.0	1.0	0.0	0.0	2.0	0.5	..	..
444	93.5	4.0	0.0	0.0	1.0	0.0	0.0	1.5	..	..	..
450*	66.0	0.0	32.0	1.5	0.0	0.5	0.0	0.0	1.0	37	63
451	53.5	0.0	45.5	0.0	0.0	0.0	0.5	0.5	1.5	60	40
462	58.5	0.5	39.0	0.5	0.0	1.0	0.0	0.5	2.0	60	40
477	64.0	22.5	2.5	0.5	1.0	0.0	4.5	5.0	11.0	0	16
506	92.0	6.0	0.0	0.0	0.0	0.0	2.0	0.0	14.0	0	15
507	75.5	15.0	0.0	2.0	0.0	1.0	6.0	0.5	10.5	2	98
514	48.5	35.5	3.0	0.0	2.5	0.0	10.0	0.5	10.0	0	12
515*	76.0	9.5	8.5	1.5	2.0	0.0	0.5	2.0	5.0	1	6
517	62.0	25.5	0.5	0.0	2.5	0.0	9.5	0.0	12.5	4	96
522	82.5	12.0	1.0	1.0	0.5	0.5	2.5	0.0	16.0	..	..
523	70.0	13.5	0.0	0.0	1.0	0.5	12.5	2.5	14.0	9	91
526	83.5	3.5	0.0	0.0	1.0	1.0	10.5	0.5	22.0	3	97
527	82.0	12.5	0.5	0.0	0.5	0.0	3.5	1.0	12.5	4	96
530*	77.5	11.5	0.0	0.5	0.5	0.5	4.5	5.0	10.5	24	76
532*	55.5	4.0	32.5	0.5	0.0	0.0	3.5	4.0	13.0	0	22
533*	69.5	8.0	4.5	2.5	0.0	0.0	10.0	5.5	12.0	..	..
535*	82.0	4.0	7.0	1.0	0.0	0.0	6.0	0.0	12.5	1	11
					Devonian						
345*	70.5	0.0	29.5	0.0	0.0	0.0	0.0	0.5	1.0	70	6
346*	53.5	0.0	46.0	0.0	0.0	0.0	0.0	0.0	0.0	86	11

Sample
347*
348
349
350
351
421
457*
458*
503
531*
534*
537*
352
353*†
354
356
357
358
359*
360*
362*
363*
364*
422*
430
431*†
435
455*
456*
459*†
461
464*†
467
471
520*
521*
528*
536*
538*
540*
542*
331
332
333
334*†
335
336
337
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341*
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343
344
355
365

APPENDIX 2. *Continued*

Sample	Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Rock fragments	Miscellaneous	Polycrystalline quartz	Tourmaline Round	Tourmaline Angular
347*	95.5	0.0	4.5	0.0	0.0	0.0	0.0	0.0	0.0	111	10
348	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.5	50	1
349	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.5	..	..
350	86.0	0.0	24.0	0.0	0.0	0.0	0.0	0.0	1.5	88	21
351	71.0	0.0	29.0	0.0	0.0	0.0	0.0	0.0	0.0	..	..
421	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	1.0	24	4
457*	74.0	0.0	26.0	0.0	0.0	0.0	0.0	0.0	0.0	39	3
458*	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	5	0
503	99.0	0.0	1.0	0.0	0.0	0.0	0.0	0.0	0.0	7	0
531*	99.0	0.0	1.0	0.0	0.0	0.0	0.0	0.0	0.0	..	..
534*	71.5	0.0	26.0	0.0	0.0	0.0	0.0	2.5	0.0	..	..
537*	68.0	0.0	32.0	0.0	0.0	0.0	0.0	0.0	0.0	17	1
					Ordovician						
352	99.5	0.0	0.0	0.0	0.0	0.0	0.0	0.5	0.0	31	1
353*††	93.5	6.5	0.0	0.0	0.0	0.0	0.0	0.0	0.5	89	11
354	98.5	1.5	0.0	0.0	0.0	0.0	0.0	0.0	0.0	99	1
356	96.5	3.0	0.0	0.0	0.5	0.0	0.0	0.0	2.0	14	15
357	99.5	0.5	0.0	0.0	0.0	0.0	0.0	0.0	2.0	92	8
358	93.0	0.5	3.0	3.5	0.0	0.0	0.0	0.0	2.0	97	3
359*	75.5	0.0	23.0	1.0	0.0	0.0	0.0	0.5	0.0	96	4
360*	99.5	0.0	0.5	0.0	0.0	0.0	0.0	0.0	0.5	..	..
362*	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	..	107	5
363*	99.0	1.0	0.0	0.0	0.0	0.0	0.0	0.0	0.5	96	4
364*	98.5	1.5	0.0	0.0	0.0	0.0	0.0	0.0	0.5	6	0
422*	70.5	0.0	29.5	0.0	0.0	0.0	0.0	0.0	0.5	95	5
430	96.0	2.0	1.5	0.0	0.0	0.0	0.0	0.0	0.5	95	5
431*††	6.0	0.0	92.0	0.0	0.0	0.0	0.0	2.0	1.0	20	0
435	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	92	8
455*	96.0	0.0	4.0	0.0	0.0	0.0	0.0	0.0	0.5	..	..
456*	84.5	0.0	14.5	1.0	0.0	0.0	0.0	0.0	0.5	46	3
459*††	42.5	0.0	56.5	1.0	0.0	0.0	0.0	0.0	1.0	11	2
461	99.5	0.0	0.0	0.5	0.0	0.0	0.0	0.0	0.0	97	3
464*††	34.0	0.0	65.5	0.5	0.0	0.0	0.0	0.0	1.5	..	..
467	77.5	14.0	0.0	3.5	0.0	0.0	1.5	3.5	1.0	33	1
471	94.0	4.0	0.0	0.0	0.0	0.0	0.0	2.0	0.0	100	0
520*	96.0	3.0	0.0	1.0	0.0	0.0	0.0	0.0	1.0	..	..
521*	99.0	0.5	0.0	0.5	0.0	0.0	0.0	0.0	0.5	..	..
528*	70.0	0.0	30.0	0.0	0.0	0.0	0.0	0.0	0.0	..	..
536*	92.0	8.0	0.0	0.0	0.0	0.0	0.0	0.0	0.5	24	1
538*	99.5	0.5	0.0	0.0	0.0	0.0	0.0	0.0	0.0	..	..
540*	90.5	1.5	3.5	4.5	0.0	0.0	0.0	0.0	0.5	47	53
542*	62.0	0.0	38.0	0.0	0.0	0.0	0.0	0.0	0.5	21	6
					Cambrian						
331	77.5	0.0	22.5	0.0	0.0	0.0	0.0	0.0	0.5	89	11
332	72.5	1.5	0.0	19.5	0.0	0.0	0.5	6.0	1.0	26	28
333	72.0	4.0	1.0	22.5	0.0	0.0	0.0	0.5	1.0	59	41
334*††	38.5	0.0	61.0	0.5	0.0	0.0	0.0	0.0	0.5	..	..
335	59.5	1.0	18.0	9.5	0.0	0.0	0.0	12.0	1.0	..	..
336	73.5	3.0	11.0	12.5	0.0	0.0	0.0	0.0	0.5	88	25
337	96.5	3.5	0.0	0.0	0.0	0.0	0.0	0.0	2.5	..	..
338	97.5	0.5	0.0	0.5	0.0	0.0	0.0	1.5	0.0	..	..
339	50.5	4.0	1.0	40.5	1.0	0.0	0.0	3.0	..	..	..
340	52.5	6.0	0.0	33.0	2.5	0.0	0.0	6.0	0.0	..	..
341*	62.5	2.5	0.0	1.0	0.0	0.5	0.0	3.5	1.5	105	2
342	97.5	2.5	0.0	0.0	0.0	0.0	0.0	0.0	0.0	..	..
343	99.5	0.0	0.0	0.5	0.0	0.0	0.0	0.0	0 0	86	22
344	95.5	2.0	0.0	1.0	0.0	0.0	0.0	1.5	1.0	..	..
355	100.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.5	81	5
365	95.0	0.0	0.0	5.0	0.0	0.0	0.0	0.0	0.5	..	..

APPENDIX 2. *Continued*

Sample	Quartz	Matrix	Carbonate	Feldspar	Mica	Chert	Rock fragments	Miscellaneous	Polycrystalline quartz	Tourmaline Round	Tourmaline Angular
366	79.5	0.0	0.0	19.5	0.0	0.0	0.0	1.0	0.0	..	..
425*††	18.5	0.0	77.0	0.5	0.0	2.5	0.0	1.5	..	..	..
433	67.0	7.0	0.0	22.0	0.5	0.0	0.0	3.0	0.0	..	..
434	87.5	10.5	0.0	1.5	0.0	0.0	0.0	0.5	3.0	75	25
436	49.5	0.0	31.5	11.5	0.0	0.0	0.0	7.5	1.0	27	0
438											
439	99.0	0.0	0.0	0.0	0.0	0.0	0.0	1.0	4.5	..	..
440	97.0	1.5	0.0	1.5	0.0	0.0	0.0	0.0	3.0	..	..
441	89.0	1.0	0.0	8.0	0.0	2.0	0.0	0.0	2.0	..	..
442	92.0	21.0	0.0	3.0	0.0	1.0	0.0	0.0	1.0	..	..
453	35.0	1.0	40.5	13.5	0.5	0.0	0.0	9.5	1.0	..	..
454	82.0	8.5	0.0	1.5	0.0	8.0	0.0	0.0	0.5	..	..
466*††	29.0	2.5	54.0	6.0	0.0	0.0	0.0	8.5	0.0	..	..
468	70.5	11.0	0.0	16.5	0.5	0.0	1.0	0.5	1.0	59	41
469	84.0	1.5	0.0	3.0	0.0	0.0	0.0	11.5	1.0	46	29
470	92.5	1.5	1.5	4.0	0.0	0.0	0.5	0.0	0.5	7	3
472*††	18.0	0.5	74.5	2.5	0.0	0.0	0.0	4.5	0.5	..	..
502	60.0	7.5	0.0	13.0	0.0	0.0	0.0	19.5	0.0	84	16
504	89.0	3.5	4.0	3.5	0.0	0.0	0.0	0.0	1.5	37	9
505	96.5	0.5	0.0	0.5	0.0	0.5	1.0	1.0	11.0	23	28
508	77.5	5.5	0.0	9.5	0.5	1.0	1.5	4.5	14.5	2	8
509	93.0	0.0	0.0	6.5	0.0	0.0	0.5	0.0	5.5	38	62
510	81.0	9.0	0.5	7.5	0.0	0.0	0.0	2.0	7.0	76	24
511	61.0	30.0	0.0	7.5	1.0	0.0	0.0	0.5	1.5	61	39
512	93.0	5.0	0.0	0.0	0.0	1.0	1.0	0.0	3.0	74	26
513*††	17.0	0.0	82.5	0.5	0.0	0.0	0.0	0.0	0.5	6	1
516	61.0	13.5	0.0	18.5	0.0	5.0	2.0	0.0	15.0	..	..
518	84.0	6.5	0.0	7.5	0.5	1.5	0.0	0.0	7.0	..	..
519	91.5	4.0	0.0	2.5	0.0	0.5	1.0	0.5	5.0	53	29
525	46.5	33.0	0.0	17.0	1.5	1.5	0.0	0.5	8.0	26	74
529	75.5	14.5	9.5	0.5	0.0	0.0	0.0	0.0	3.0	43	3

*Subsurface sample
†Age and location given by Siever (1953, Table 1)
††Not included in Table 2

TOURMALINE ANGULARITY OF SAMPLES IN SYSTEMS STUDIED

Sample	Round	Angular	Sample	Round	Angular	Sample	Round	Angular	Sample	Round	Angular
	Tertiary		235	2	98	H- 25*	44	56	H- 60	40	3
B-1	4	61	256	0	103	H- 26*	53	47	H- 61	94	0
Car-2	4	21	269	4	96	H- 27*	42	58			
C-3	4	26	278	1	99	H- 33A	47	53		*Devonian*	
C-16†	7	21	279	2	75	H- 34A	37	63	H- 62*	94	0
C-16††	2	18	319	8	90	H- 35A	36	64	H- 63	17	0
C-17†	6	48	324	0	100	H- 36A	33	67	H- 64	37	2
C-17††	7	22	325	4	96	H- 37A	47	53	H- 65*	29	8
C-20	5	81	H- 92	0	99	H- 38A	49	51	H- 66*	6	0
D-1	9	18	H- 93	3	97	H- 39A	59	41	H- 90	95	5
G-37	7	41	H- 94	1	10	H- 40A	49	51			
H-5	11	50	H- 95	0	105	H- 41A	64	36		*Ordovician*	
H-9	1	25	H- 96	2	96	H- 42*	35	65	H- 67	2	0
KH-4	8	34	H- 97	1	102	H- 43	41	59	H- 68	72	28
Mass-6	22	74	H- 98	1	101	H- 44	44	56	H- 69	100	0
Mc-1	7	40	H- 99	2	101	H- 45	44	56	H- 70	101	2
M-24	4	37	H-100	2	103	H- 46	51	49	H- 91	21	0
P-2	6	10	H-101	1	107	H- 47	44	56			
P-4†	2	8	H-102	0	96	H- 48*	28	72		*Cambrian*	
P-4††	3	17				H- 49	50	50			
S-7	3	7		*Upper Mississippian*		H- 50	36	64	H- 71	40	67
Sc-2	4	22		(*Chester*)		H- 51	44	56	H- 72	68	41
TH-4	2	23				H- 52	55	45	H- 73	74	4
IE-1	4	4	H- 1	49	51	H- 53	48	52	H- 74	111	0
680	2	11	H- 2	33	67	H- 54	33	67	H- 75	105	1
			H- 3	31	69	H- 55	64	36	H- 76	86	14
	Middle and Upper		H-13*	55	45	H- 56	27	73	H- 77	26	6
	Pennsylvanian		H-14*	49	51	H- 57	41	59	H- 78*	8	11
48	3	101	H-15*	40	60	H-103	36	64	H- 79*	48	20
62	2	43	H-16*	52	48	H-104	46	54	H- 80*	51	58
63	2	97	H-17*	45	55	H-105	41	59	H- 81*	50	14
83	2	100	H-18*	63	37	H-106	49	51	H- 82*	7	0
89	0	30	H-19	55	45				H- 83*	38	3
98	2	62	H-20*	56	44				H- 84*	4	8
101	13	87	H-21*	50	50		*Middle and Lower*		H- 85*	16	27
103	14	72	H-22*	43	57		*Mississippian*		H- 86*	44	31
108	9	89	H-23*	68	32	H- 58	94	6	H- 87*	56	16
234	2	89	H-24*	57	43	H- 59	5	95	H- 88	74	14

*Subsurface sample
†Age and location given by Siever (1953, Table 1)
††Not included in Table 2

APPENDIX 2. *Continued*

LOCATION OF SAMPLES USED IN STUDY

Sample	Unit	Location
		Tertiary
B- 1	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Ballard County, Ky.
Car- 2	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Carlisle County, Ky.
C- 3	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Caldwell County, Ky.
C- 16	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Caldwell County, Ky.
C- 16	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Caldwell County, Ky.
C- 17	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Caldwell County, Ky.
C- 17	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Caldwell County, Ky.
C- 20	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Caldwell County, Ky.
D- 1	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Dunklin County, Mo.
G- 37	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Graves County, Ky.
H- 5	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Henry County, Tenn.
H- 9	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Henry County, Tenn.
KH- 4	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Hickman County, Ky.
Mass-6	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Massac County, Ill.
Mc- 1	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), McCracken County, Ky.
M- 24	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Marshall County, Ky.
P- 2	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Pulaski County, Ill.
P- 4	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Pulaski County, Ill.
P- 4	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Pulaski County, Ill.
S- 7	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Stoddard County, Mo.
Sc- 2	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Scott County, Mo.
TH- 4	Pliocene	Lafayette gravel from Smithland Surface (Potter, 1955, Fig. 17), Henry County, Tenn.
IE- 1	Eocene	Gravel pit, NW¼ NW¼ sec. 18, T. 16 S., R. 1 W., Pulaski County, Ill.
680	Eocene	Road cut on Rt. 440, 4 miles east of Wycliffe, Ballard County, Ky.
681	Eocene	Clay pit, sec. 27, T. 15 S., R. 1 E., Pulaski County, Ill.
682	Eocene	Stream cut, sec. 9, T. 26 N., R. 11 E., Stoddard County, Mo.
683	Eocene	Railroad cut on Highway 57, Hardeman County, Tenn.
684	Eocene	Road cut, 2 miles west of Pureyear, Henry County, Tenn.
685	Pliocene	Gravel pit, sec. 13, T. 15 S., R. 1 E., Pulaski County, Ill.
		Middle and Upper Pennsylvanian
48	McLeansboro	NW¼ NE¼ SE¼ sec. 16, T. 8 S., R. 3 E., Williamson County, Ill.
62	Curlew	SW cor. SE¼ SW¼ sec. 22, T. 10 S., R. 6 E., Saline County, Ill.
63	Murray Bluff	NW cor. NE¼ SW¼ sec. 27, T. 10 S., R. 6 E., Saline County, Ill.
83	Isabel	Near Marietta Station, sec. 21, T. 6 N., R. 1 E., Fulton County, Ill.
89	Brazil	SW¼ sec. 27, T. 16 N., R. 7 W., Parke County, Ind.
98	Curlew	Green River at Morgantown, Butler County, Ky.
101	Sebree	1.5 miles west of Sebree, Webster County, Ky.
103	Aberdeen	18,300 F.W.L.; 2000 F.N.L.; Morgantown quadrangle, Butler County, Ky.
108	Pope Creek	Sec. 25, T. 17 N., R. 1 E., Rock Island County, Ill.
234	Henshaw	Dixon, Webster County, Ky.
235	Mt. Gilead	Roadside exposure just south of former Mt. Gilead School, near Slaughters, Webster County, Ky.
256	McLeansboro	Sandstone 8 feet below Scottville limestone SW¼ sec. 16, T. 12 N., R. 9 W., Macoupin County, Ill.
269	Lisman	Three-eighths of a mile south of Coiltown, Hopkins County, Ky.
278	Merom	At Merom Bluff, SW¼ SW¼ sec. 7, T. 1 N., R. 10 W., Sullivan County, Ind.
279	Lisman	Linton No. 28 strip mine, SE¼ SE¼ sec. 15, T. 8 N., R. 7 W., Greene County, Ind.
319	Delwood	SW¼ SE¼ NE¼ sec. 19, T. 9 S., R. 2 W., Jackson County, Ill.
324	Stonefort	Madison Coal Co. Hole No. 3, sec. 29, T. 13 N., R. 5 W., Christian County, Ill.
325	Gimlet	SE¼ NE¼ SW¼ sec. 29, T. 30 N., R. 4 E., Livingston County, Ill.
H- 92	McLeansboro	Approximately 30–50 feet below Livingston limestone, sec. 26, T. 19 N., R. 13 W., Vermilion County, Ill.
H- 93	Pleasantview	SW¼ SW¼ SW¼ sec. 24, T. 12 N., R. 11 W., Greene County, Ill.
H- 94	Vermilionville	18 feet below coal No. 11, Webster County, Ky.
H- 95	Anvil Rock	Sec. 8, T. 10 S., R. 9 E., Gallatin County, Ill.
H- 96	McLeansboro	Above Lonsdale limestone in sec. 5, T. 19 N., R. 12 W., Vermilion County, Ill.
H- 97	Mt. Carmel	Sec. 21, T. 1 S., R. 12 W., White County, Ill.
H- 98	Gimlet	SE cor. sec. 25, T. 25 N., R. 7 E., Peoria County, Ill.
H- 99	Palzo	Superior No. 2 Mussett, sec. 1, T. 3 S., R. 10 E., Edwards County, Ill.
H-100	Palzo	NW¼ SW¼ NE¼ sec. 30, T. 10 S., R. 6 E., Saline County, Ill.
H-101	McLeansboro	Sandstone approximatley at the horizon of the Livingstone limestone SE¼ NE¼ SE¼ sec. 36, T. 8 N., R. 14 W., Crawford County, Ill.

APPENDIX 2. *Continued*

Sample	Unit	Location
H-102	McLeansboro	Sandstones at position of Greenup limestone, sec. 19, T. 9 N., R. 8 E., Cumberland County, Ill.
		Upper Mississippian (Chester)
368	Aux Vases	Sec. 11, T. 34 N., R. 13 E., Perry County, Mo.
370	Hardinsburg	1 mile north of Hardinsburg, Breckinridge County, Ky.
371	Cypress	NE¼ NE¼ sec. 14, T. 14 S., R. 6 E., Pope County, Ill.
372	Bethel	Sec. 30, T. 13 S., R. 6 E., Pope County, Ill.
373	Bethel	SE¼ SE¼ sec. 29, T. 6 N., R. 2 W., Lawrence County, Ind.
374	Bethel	Top of the Main Cedar Bluff quarry, 2.5 miles south of Princeton, Caldwell County, Ky.
426	Cypress	Shell Oil Co., Montgomery No. 1, sec. 14, T. 1 S., R. 6 E., Wayne County, Ill., 3077 feet
427	Cypress	Pure Oil Co., Harrell No. 1, sec. 26, T. 4 N., R. 9 E., Richland County, Ill., 2576 feet
429	Yankeetown (Benoist)	Townsend-Clark No. 1, SE¼ NE¼ SE¼ sec. 13, T. 2 S., R. 1 E., Jefferson County, Ill., 2090 feet
448	Bethel (Mooretown)	Cut on south side of U. S. Highway 60 approximately ½ mile south of Guston, sec. 7-Q-40, Meade County, Ky.
465	Hardinsburg-Big Clifty	NW¼ sec. 11, T. 5 S., R. 1 W., Perry County, Ind.
473	Hardinsburg	From ledge 3 feet above base of formation in abandoned quarry on north side of Green River at Brownsville, Edmonson County, Ky.
474	Tar Springs	Sandefur, Engle No. 3, sec. 14-N-24, Webster County, Ky.
475	Waltersburg	AMS Oil Co., Austin No. 1, sec. 28, T. 6 S , R. 9 E., White County, Ill.
476	Tar Springs	2 miles west of Cloverport, Breckinridge County, Ky.
479	Palestine	SW¼ SE¼ NW¼ sec. 11, T. 3 S., R. 3 W., DuBois County, Ind.
480	Aux Vases	NW¼ SW¼ NE¼ sec. 21, T. 9 N., R. 2 W., Monroe County, Ind.
481	Elwren (Cypress)	SW¼ SW¼ NE¼ sec. 20, T. 8 N., R. 2 W., Monroe County, Ind.
482	Elwren (Cypress)	SE¼ NE¼ NW¼ sec. 32, T. 2 S., R. 1 E., Crawford County, Ind.
483	Waltersburg	NE¼ NE¼ sec. 22, T. 3 S., R. 3 W., Perry County, Ind.
484	Sample	NE¼ NW¼ NE¼ sec. 12, T. 3 S., R. 2 E., Harrison County, Ind.
485	Big Clifty	SE¼ SE¼ NW¼ sec. 6, T. 4 S., R. 2 E., Crawford County, Ind.
487	Palestine	8430 F.W.L.; 13,800 F.S.L., I-15, Livingston County, Ky.
573	Elwren (Cypress)	5000 F.W.L.; 7700 F.N.L., sec. 6-P-39, Breckinridge County, Ky.
574	Big Clifty	11,500 F.E.L.; 3500 F.S.L., sec. 23-K-41, Hart County, Ky.
575	Hardinsburg	10,500 F.S.L.; 7050 F.E.L., L-41, Grayson County, Ky.
577	Sample	5000 F.N.L.; 7700 F.N.L., sec. 6-P-39, Breckinridge County, Ky.
581	Bethel	Texas Co., J. P. McGurie No. 1, sec. 15, T. 9 S., R. 8 E., Gallatin County, Ill., 2366 feet
582	Aux Vases	Webster and Shirk, N. Zipk No. 2, sec. 35, T. 5 N., R. 6 E., Clay County, Ill., 2731 feet
583	Aux Vases	Doran, A. W. Kamp No. B-4, sec. 30, T. 5 N., R. 3 E., Fayette County, Ill., 1919 feet
584	Aux Vases	Superior Oil, Benton *et al.* No. 1, sec. 25, T. 4 S., R. 4 E., Jefferson County, Ill., 3229 feet
585	Sample (Paint Creek)	Mitchell, Lamont No. 1, sec. 4, T. 4 S., R. 10 E., White County, Ill., 3003 feet
586	Aux Vases	Branyan, Laflin No. 1, NW¼ sec. 15, T. 11 N., R. 4 E., Shelby County, Ill., 3347, 1906 feet
587	Bethel	LaGrange, Huffman No. 1, sec. 4, T. 3 N., R. 1 W., Clinton County, Ill., 1466 feet
588	Bethel	Duncan, Barbre No. 1, sec. 25, T. 4 S., R. 9 E., White County, Ill., 2990 feet
H- 1	Degonia	SW¼ sec. 33, T. 7 S., R. 6 W., Randolph County, Ill.
H- 2	Degonia	Sec. 20, T. 11 S., R. 1 W., Union County, Ill.
H- 3	Degonia	Along line of secs. 10 and 11, T. 5 S., R. 1 W., Perry County, Ind.
H- 13	Clore	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 14	Clore	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 15	Clore	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 16	Clore	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 17	Palestine	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 18	Tar Springs	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 19	Tar Springs	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 20	Tar Springs	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 21	Tar Springs	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 22	Hardinsburg	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 23	Cypress	J. H. Forester, J. H. Forester No. 1, sec. 5, T 6 S., R. 1 W., Perry County, Ill.

APPENDIX 2. *Continued*

Sample	Unit	Location
H- 24	Cypress	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 25	Paint Creek	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 26	Yankeetown	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 27	Renault	J. H. Forester, J. H. Forester No. 1, sec. 5, T. 6 S., R. 1 W., Perry County, Ill.
H- 33A	Degonia	NW¼ NW¼ NW¼ sec. 18, T. 11 S., R. 9 E., Hardin County, Ill.
H- 34A	Palestine	NE¼ NW¼ NW¼ sec. 18, T. 11 S., R. 9 E., Hardin County, Ill.
H- 35A	Palestine	NE¼ NW¼ NW¼ sec. 24, T. 11 S., R. 9 E., Hardin County, Ill.
H- 36A	Waltersburg	SE¼ SW¼ SE¼ sec. 19, T. 11 S., R. 10 E., Hardin County, Ill.
H- 37A	Tar Springs	SW¼ SE¼ sec. 27, T. 11 S., R. 10 E., Hardin County, Ill.
H- 38A	Tar Springs	25-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
H- 39A	Tar Springs	SW¼ SE¼ sec. 27, T. 11 S., R. 10 E., Hardin County, Ill.
H- 40A	Hardinsburg	151-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
H- 41A	Hardinsburg	222-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
H- 42	Tar Springs	Oil well core at 2098 feet, sec. 36, T. 6 S., R. 2 E., Franklin County, Ill.
H- 43	Tar Springs	3.7 miles south of Crofton, Christian County, Ky., on U. S. 41
H- 44	Hardinsburg	3400 F.W.L.; 1700 F.N.L., sec. 5-H-42, Edmonson County, Ky.
H- 45	Hardinsburg	3800 F.N.L.; 1950 F.W.L., sec. 5-R-38, Meade County, Ky.
H- 46	Hardinsburg	550 F.S.L.; 400 F.E.L., sec. 11-S-37, Meade County, Ky.
H- 47	Hardinsburg	Sec. 6, T. 7 S., R. 2 E., Crawford County, Ind.
H- 48	Hardinsburg	Deep Rock, Phipps No. 1, sec. 29, T. 5 S., R. 9 E., White County, Ill., 2647 feet
H- 49	Big Clifty	2800 F.E.L.; 9100 F.N.L., sec. 10-O-39, Breckinridge County, Ky.
H- 50	Big Clifty	Sec. 31, T. 3 S., R. 1 E., Crawford County, Ind.
H- 51	Big Clifty	Approximately 3.5 miles north of Fairview in secs 8 or 9-E-27, Todd County, Ky.
H- 52	Big Clifty	3000 F.E.L.; 8700 F.S.L., sec. 20-M-40, Grayson County, Ky.
H- 53	Big Clifty	5350 F.E.L.; 10,900 F.W.L., I-44, Hart County, Ky.
H- 54	Elwren (Cypress)	Sec. 21, T. 13 N., R. 5 W., Putnam County, Ind.
H- 55	Sample	8150 F.W.L.; 8300 F.N.L., sec. 7, L-42, Grayson County, Ky.
H- 56	Bethel	Sec. 3, T. 3 N., R. 2 W., Lawrence County, Ind.
H- 57	Cypress	Sec. 20, T. 13 S., R. 2 E., Johnson County, Ill.
H-103	Cypress	426-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
H-104	Cypress	463-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
H-105	Paint Creek	484-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
H-106	Bethel	515-foot diamond-drill hole, NE¼ sec. 27, T. 11 S., R. 9 E., Hardin County, Ill.
		Middle and Lower Mississippian
369	Spar Mountain	Direction, Strong No. 1, sec. 22, T. 12 N., R. 7 E., Coles County, Ill., 1973 feet
420	Ste. Genevieve	Type Hoffner section in SW¼ NE¼ SW¼ sec. 8, T. 13 S., R. 1 E., Union County, Ill.
424	Bushberg	SW¼ SW¼ SE¼ sec. 15, T. 44 N., R. 4 E., St. Louis County, Mo.
428	Borden (Carper)	Rudy, Ochs No. 1, NW¼ NE¼ NW¼ sec. 23, T. 6 N., R. 10 E., Jasper County, Ill., 4126 feet
432	Sylamore	At Pinnacle Hill in NW¼ sec. 16, T. 53 N., R. 1 E., Pike County, Mo.
444	Bushberg	NE¼ SE¼ sec. 53, T. 46 N., R. 10 W., Calloway County, Mo.
450	Ste. Genevieve	Howard and Howard, Leslie No. 1, SE¼ NE¼ SE¼ sec. 22, T. 5 S., R. 5 E., Hamilton County, Ill., 3386–3392 feet
451	St. Louis or Salem	Near Keokuk, Lee County, Iowa
462	St. Louis	Cooks quarry, near Ames, Story County, Iowa
477	Harrodsburg	NW¼ NW¼ sec. 35, T. 2 N., R. 4 E., Washington County, Ind.
506	Marshall	Sec. 21, T. 19 N., R. 15 E., Huron County, Mich.
507	Napoleon	Sec. 31, T. 3 S., R. 2 E., Jackson County, Mich.
514	Marshall	Along shore of Lake Huron, sec. 23, T. 19 N., R. 13 E., Huron County, Mich.
515	Berea	Ohio, Cooper No. 1, SW¼ SW¼ SE¼ sec. 7, T. 1 S., R. 3 E., Washtenaw County, Mich.
517	Marshall	Sec. 15, T. 2 S., R. 7 W., Calhoun County, Mich.
522	Marshall	NW¼ NW¼ sec. 13, T. 14 N., R. 12 E., St. Clair County, Mich.
523	Marshall	Sec. 21, T. 19 N., R. 13 E., Huron County, Mich.
526	Marshall	NW¼ sec. 31, T. 4 S., R. 2 W., Jackson County, Mich.
527	Marshall	SE¼ NE¼ NE¼ sec. 3, T. 4 S., R. 3 W., Jackson County, Mich.
530	Berea	Wickland, Frost No. 1, sec. 20, T. 10 N., R. 6 E., Saginaw County, Mich., 1532 feet
532	Meramec-Osage	Sun, Menningel-Church *et al.* No. 1, sec. 8, T. 19 N., R. 3 W., Clare County, Mich., 1409 feet
533	Meramec-Osage	Sun, Menningel-Church *et al.* No. 1, sec. 8, T. 19 N., R. 3 W., Clare County, Mich., 1421 feet
535	Marshall	Sun, Grande *et al.* No. 2, sec. 36, T. 20 N., R. 4 W., Clare County, Mich., 1625 feet

APPENDIX 2. *Continued*

Sample	Unit	Location
H- 58	Bushberg	SW¼ SW¼ SE¼ sec. 15, T. 44 N., R. 4 E., St. Louis County, Mo.
H- 59	Salem(Sonora)	SE¼ sec. 8, T. 7 N., R. 7 W., Hancock County, Ill.
H- 60	Glen Park	SW¼ sec. 17, T. 6 S., R. 5 W., Pike County, Ill.
H- 61	Sylamore	SE¼ sec. 7, T. 7 S., R. 4 W., Pike County, Ill.
		Devonian
345	Cedar Valley	Obering, Dellinger No. 1, sec. 15, T. 14 N., R. 4 E., Moultrie County, Ill.
346	Lingle (Devonian)	Texas, Miller No. 1, sec. 22, T. 6 N., R. 2 W., Bond County, Ill.
347	Devonian	Thompson, Randall No. 1-D, sec. 26, T. 19 N., R. 2 W., Boone County, Ind.
348	Cedar Valley	NW¼ NW¼ sec. 4, T. 8 N., R. 13 W., Jersey County, Ill.
349	Beauvaris	1½ miles south of Ozora, Ste. Genevieve County, Mo.
350	Grand Tower	NE¼ SW¼ NE¼ sec. 23, T. 10 S., R. 4 W., Jackson County, Ill.
351	Cedar Valley	SE¼ NE¼ sec. 29, T. 9 N., R. 13 W., Jersey County, Ill.
421	Grand Tower	NW¼ sec. 27, T. 11 S., R. 2 W., Union County, Ill.
457	Cedar Valley (Devonian)	Obering, Dellinger No. 1, NE¼ NW¼ SE¼ sec. 15, T. 14 N., R. 4 E., Moultrie County, Ill., 2745–2748 feet
458	Lingle (Devonian)	California, Kunz No. 1, SW¼ SE¼ SE¼ sec. 1, T. 3 N., R. 6 W., Madison County, Ill., 1884–1889 feet
503	Sylvania	Sec. 14, T. 3 S., R. 10 E., Wayne County, Mich.
531	Sylvania	Mogal, Domokos No. B-1, sec. 11, T. 18 N., R. 1 W., Gladwin County, Mich., 5344 feet
534	Richfield	Sun, Caldwell No. A-1, sec. 19, T. 23 N., R. 8 W., Missaukee County, Mich., 5034 feet
537	Sylvania	Dow, Dow No. 23, sec. 33, T. 14 N., R. 1 E., Midland County, Mich., 3160 feet
H- 62	Cedar Valley (Hoing)	Ohio, Binney No. 24, sec. 16, T. 4 N., R. 4 W., McDonough County, Ill.
H- 63	Cedar Valley (Hoing?)	SW¼ SE¼ sec. 15, T. 56 N., R. 5 W., Ralls County, Mo.
H- 64	Cedar Valley	Sec. 11, T. 12 S., R. 2 W., Calhoun County, Ill.
H- 65	Basal Lingle	Sun, House No. 1, sec. 32, T. 10 N., R. 3 W., Gratiot County, Mich., 2057 feet
H- 66	Sylvania	Dow, Dow No. 23, sec. 33, T. 14 W., R. 1 E., Midland County, Mich.
H- 90	Dutch Creek	Along Ill. State Route 127 in NW¼ NE¼ sec. 15, T. 12 S., R. 2 W., Union County, Ill.
		Ordovician
352	St. Peter	On Wisconsin State Route 92 north of Mt. Vernon, Dane County, Wis.
353	St. Peter	Superior, Ford No. C-17, sec. 27, T. 4 S., R. 14 W., White County, Ill., 7490 feet
354	St. Peter	NE¼ NW¼ SE¼ sec. 31, T. 40 N., R. 6 E., Jefferson County, Mo.
356	Gunter	SW¼ NW¼ sec. 18, T. 27 N., R. 1 E., Carter County, Mo.
357	New Richmond	Sec. 6, T. 35 N., R. 5 E., LaSalle County, Ill.
358	Oneota	NW¼ SE¼ NW¼ sec. 21, T. 7 N., R. 9 E., Dane County, Wis.
359	Glenwood	Carter, Seaman No. 2, sec. 2, T. 11 N., R. 7 E., Coles County, Ill., 4691–4695 feet
360	St. Peter	Mississippi River Fuel, Theobold No. A-15, sec. 35, T. 1 S., R. 10 W., Monroe County, Ill.
362	St. Peter	SE¼ SE¼ SE¼ sec. 19, T. 6 N., R. 2 W., Calhoun County, Ill.
363	St. Peter	Texas, Pat No. 3?, sec. 5, T. 1 N., R. 2 E., Marion County, Ill.
364	St. Peter	Sun, Dameny No. 1, sec. 5, T. 15 N., R. 1 E., Macon County, Ill.
422	Uppermost Knox	Superior, Ford No. C-17, sec. 22, T. 4 S., R. 14 W., White County, Ill., 7677 feet
430	New Richmond	2 miles west of junction of U. S. 52 and U. S. 16 near Preston, T. 103 N., R. 11 W., Fillmore County, Minn.
431	New Richmond	Center of W½ SE¼ SW¼ sec. 10, T. 7 N., R. 7 E., Dane County, Wis.
435	Kasota	Sec. 32, T. 110 N., R. 26 W., Le Seur County, Minn.
455	Cotter	Mississippi River Fuel, Theobald No. A-5, sec. 35, T. 1 S., R. 10 W., Monroe County, Ill.
456	Powell	Mississippi River Fuel, Theobald No. A-5, sec. 35, T. 1 S., R. 10 W., Monroe County, Ill.
459	Roubidoux	East side of Missouri Highway 7, E/2 NW¼ sec. 36, T. 28 N., R. 7 W., Texas County, Mo.
461	St. Peter	SW¼ sec. 2, T. 94 N., R. 3 W., Clayton County, Iowa
464	Oneota	Sec. 6, T. 108 N., R. 26 W., Blue Earth County, Minn.
467	Thebes	NW¼ SW¼ SW¼ sec. 9, T. 15 S., R. 3 W., Alexander County, Ill.
471	St. Peter	NE¼ sec. 35, T. 95 N., R. 3 W., Clayton County, Iowa
520	Prairie du Chien?	Superior, Sippy No. 17, sec. 25, T. 17 N., R. 16 W., Mason County, Mich., 6145–6155 feet
521	Basal Ordoviican?	Superior, Sippy No. 17, sec. 25, T. 17 N., R. 16 W., Mason County, Mich., 6270–6280 feet
528	Basal Trenton	Rayborn, Watchhorn and Wells No. 1, sec. 5, T. 10 N., R. 9 E., Tuscola County, Mich.

APPENDIX 2. *Continued*

Sample	Unit	Location
536	St. Peter	Superior, Sippy No. 17, sec. 25, T. 17 N., R. 16 W., Mason County, Mich., 6037 feet
538	St. Peter	Sun, Bradley No. 4, sec. 11, T. 12 N., R. 13 W., Newaygo County, Mich., 6518 feet
540	St. Peter	Ohio, Rehninhardt No. 1, sec. 35, T. 22 N., R. 2 E., Ogemaw County, Mich., 10,49[illegible] feet
542	St. Peter	Carter, Launer No. 2, sec. 6, T. 16 N., R. 17 W., Oceania County, Mich., 5244 feet
H- 67	New Richmond	Sec. 6, T. 108 N., R. 26 W., Blue Earth County, Minn.
H- 68	Glenwood	NE¼ SE¼ NW¼ sec. 36, T. 23 N., R. 6 E., Ogle County, Ill.
H- 69	St. Peter	NE¼ NE¼ SE¼ sec. 1, T. 23 N., R. 9 E., Ogle County, Ill.
H- 70	St. Peter	Near top of Mount Victory Hill, sec. 4, T. 23 N., R. 10 E., Ogle County, Ill.
H- 91	New Richmond	Center sec. 1, T. 27 N., R. 19 W., Pierce County, Wis.
		Cambrian
331	Jordan	South Madison railroad cut NW¼ SE¼ NE¼ sec. 26, T. 7 N., R. 9 E., Dane County, Wis.
332	Bad Axe	At Mynard Pass on Wisconsin State Route 16 near Tomah, Monroe County, Wis.
333	Mt. Simon	Near Neillsville, T. 24 N., R. 2 W., Clarke County, Wis.
334	Hudson	At Mazomanie, T. 8 N., R. 6 E., Dane County, Wis.
335	Bad Axe	At Mazomanie, T. 8 N., R. 6 E., Dane County, Wis.
336	Eau Claire	Southwest of Merrilin, T. 23 N., R. 4 W., Jackson County, Wis.
337	Galesville	At Friendship Mound, Adams County, Wis., T. 17 N., R. 6 E.
338	Ironton	At Twin Bluffs, Mauston, T. 15 N., R. 3 E., Juneau County. Wis.
339	Goodenough	Northeast of Wheeler, Dunn County, Wis., T. 30 N., R. 13 W.
340	Goodenough	At Dallas, Barron County, Wis., T. 32 N., R. 12 W.
341	Ironton	Bays *et al.*, National Gas Storage Co., NW¼ NE¼ NE¼ sec. 32, T. 30 N., R. 10 E., Kankakee County, Ill., 1742–1743 feet
342	Dresbach	At Rock Springs, Sauk County, Wis.
343	Lamotte	Sec. 2, T. 36 N., R. 6 E., Ste. Genevieve County, Mo.
344	Dresbach	NE¼ sec. 14, T. 11 N., R. 6 E., Sauk County, Wis.
355	Madison	Black Earth, Dane County, Wis., T. 8 N., R. 6 E.
365	Mt. Simon	Along Chippawa River, sec. 6, T. 27 N., R. 9 W., Eau Claire, Eau Claire County, Wis.
366	Goodenough	Opposite Camp Grounds at Taylors Falls, Minn.
425	Lamotte	SW¼ SW¼ sec. 31, T. 35 N., R. 4 E., St. Francis County, Mo.
433	Trempealeau	SW¼ NE¼ SE¼ sec. 29, T. 99 N., R. 3 W., Allamakee County, Iowa
434	Franconia	SW¼ NE¼ SE¼ sec. 29, T. 99 N., R. 3 W., Allamakee County, Iowa
436	Jordan	SE¼ sec. 15, T. 30 N., R. 20 W., Washington County, Minn.
439	Mt. Simon	NW¼ SE¼ sec. 31, T. 29 N., R. 8 W., Chippewa County, Wis. (at Precambrian contact)
440	Mt. Simon	5 feet from Precambrian contact in NW¼ SE¼ sec. 31, T. 29 N., R. 8 W., Chippewa County, Wis.
441	Mt. Simon	10 feet from Precambrian contact in sec. 31, T. 29 N., R. 8 W., Chippewa County, Wis.
442	Mt. Simon	15 feet from Precambrian contact in same location
453	Davis	E½ NW¼ NE¼ sec. 27, T. 39 N., R. 3 E., Washington County, Mo.
454	Lamotte	Ozark Ore Mine, SW¼ sec. 31, T. 35 N., R. 4 E., St. Francis County, Mo.
466	St. Lawrence	NE¼ sec. 28, T. 114 N., R. 23 W., Scott County, Minn.
468	Galesville	At Hudson, T. 29 N., R. 20 W., St. Croix County, Wis.
469	Woodhill	SW¼ sec. 36, T. 107 N., R. 8 W., Winona County, Minn.
470	Upper Mazomanie	Cen.-S½ sec. 1, T. 106 N., R. 7 W., Winona County, Minn.
472	Woodhill	Interstate Park, Taylors Falls, Chisago County, Minn.
502	Franconia	Allamakee County, Iowa.
504	Franconia	Sec. 28, T. 48 N., R. 17 W., Alger County, Mich.
505	Dresbach	Sec. 13, T. 46 N., R. 20 W., Alger County, Mich.
508	Jacobsville	Sec. 25, T. 48 N., R. 2 E., Chippewa County, Mich.
509	Dresbach	Sec. 24, T. 51 N., R. 36 W., Houghton County, Mich.
510	Franconia	Sec. 17, T. 46 N., R. 22 W., Alger County, Mich.
511	Franconia	Sec. 35, T. 43 N., R. 25 W., Delta County, Mich.
512	Dresbach	Sec. 28, T. 48 N., R. 17 W., Alger County, Mich.
513	Franconia	Sec. 27, T. 40 N., R. 30 W., Dickenson County, Mich.
516	Jacobsville	Sec. 32, T. 48 N., R. 19 W., Alger County, Mich.
518	Jacobsville	Sec. 31, T. 50 N., R. 39 W., Ontanogon County, Mich.
519	Jacobsville	Sec. 6, T. 55 N., R. 32 W., Houghton County, Mich.
525	Jacobsville	SE¼ NE¼ sec. 12, T. 54 N., R. 34 W., Houghton County, Mich.
529	Dresbach	Encomment D'Ours Island, 4600 yards F.W.L. and 4950 yards F.S.L. of the Bruce Mines 41, J15 W½ quadrangle, Ont.
H- 71	Eau Claire	At Blair, T. 21 N., R. 7 W., Trempealeau County, Wis.

APPENDIX 2. *Continued*

Sample	Unit	Location
H- 72	Mt. Simon	Mt. Simon, Eau Claire County, Wis.
H- 73	Dresbach	Friendship Mound, Adams County, Wis., T. 17 N., R. 6 E.
H- 74	Madison	Stoddard, Vernon County, Wis., sec. 27, T. 14 N., R. 7 W.
H- 75	Dresbach	Sec. 21, T. 18 N., R. 4 W., Trempealeau County, Wis.
H- 76	Mt. Simon	Eau Claire, Eau Claire County, Wis.
H- 77	Jordan	SE¼ sec. 19, T. 114 N., R. 23 W., Scott County, Minn.
H- 78	Cambrian (composite)	Tecumseh, Gibson No. 1, sec. 33, T. 29 N., R. 12 E., Allen County, Ind., 3095–3145 feet
H- 79	Cambrian	Ohio, Shaw No. 1, sec. 16, T. 36 N., R. 8 E., Douglas County, Ill., 3100–4100 feet
H- 80	Franconia	Varner, Stockton No. 1, sec. 11, T. 27 N., R. 4 E., Jo Daviess County, Ill., 910–955 feet
H- 81	Galesville	Neely and Schimmelpfenig, Warner No. 4, sec. 21, T. 35 N., R. 14 E., Cook County, Ill., 1700–1734 feet
H- 82	Mt. Simon	Amboy, McElroy No. 1, sec. 30, T. 20 N., R. 10 E., Lee County, Ill., 1925–2016 feet
H- 83	Galesville	Gray-Milaeger, Simpson No. 1, sec. 33, T. 46 N., R. 11 E., Lake County, Ill., 1155–1258 feet
H- 84	Mt. Simon	Bank of Rensselaer, Egar Estate No. 1, sec. 14, T. 30 N., R. 6 W., Jasper County, Ind., 2960–3304 feet
H- 85	Mt. Simon	Seele, Seele No. 1, sec. 24, T. 44 N., R. 2 E., Winnebago County, Ill.
H- 86	Mt. Simon	Eel River, Blacker No. 1, sec. 12, T. 30 N., R. 6 E., Kosciusko County, Ind., 2865–3295 feet
H- 87	Cambrian (composite)	Miller, Glidden Co. No. 1, sec. 33, T. 40 N., R. 13 E., Cook County, Ill., 1900–2000 feet
H- 88	Franconia	Black Earth, Dane County, Wis.

APPENDIX 3. PRESENT STREAM PATTERN OF THE UPPER MISSISSIPPI VALLEY AND ADJACENT AREAS

Within the portion of the upper Mississippi Valley and adjacent areas where our petrographic samples were obtained (Fig. 3) we wished to estimate the orientation of the higher-order streams. For this purpose we used base maps (scale 1:500,000) of the appropriate states that show the stream pattern. The direction of the stream was specified by the orientation of its tangent at the sample point.

Sample points were obtained by selecting the southeasternmost township within the county and using as the tangent point the place where the stream crossed the south side of the county line. If the southeasternmost township had no stream, successive western townships of the southern tier were examined until a stream was encountered. In those counties that did not have a unique southeastern township, the southernmost one was selected. In portions of Kentucky and Tennessee, Carter townships were used in lieu of ordinary townships.

The sampling procedure based on existing county and township boundaries was easy to use and yielded the pattern of sampling points shown in Figure 17. The orientations of 644 stream tangents had a mean value of 188°.

Short Notes

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HORIZONTAL DISPLACEMENTS IN THE FLOOR OF THE NORTHEASTERN PACIFIC OCEAN

Abstract: A total magnetic intensity survey in the northeastern Pacific Ocean revealed a north-south pattern of magnetic anomalies, which is cut through by the Murray, the Pioneer, and the Mendocino faults. The amount of slip along these faults is measured by fitting the magnetic anomaly pattern across the faults. The combined left-lateral displacement across the Mendocino and the Pioneer faults is 1420 km.

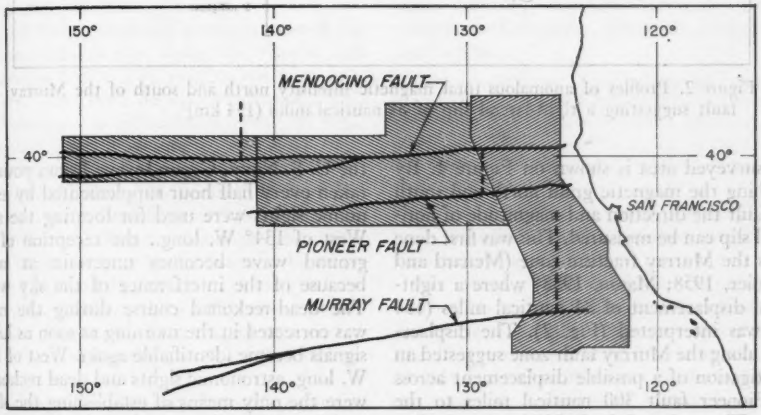

Figure 1. Location of the magnetic survey relative to the Mendocino, Pioneer, and Murray fractures. The dashed vertical line represents the axis of a north–south magnetic anomaly which was continuous before slipping started.

Displacements along strike-slip faults on land have been discussed in the literature. Kennedy (1946) has demonstrated a strike-slip displacement of 65 miles (104km) on the Great Glen fault in Scotland. According to Wellman 1952 the Alpine fault in New Zealand may have slipped 300 miles (480 km). The work of Hill and Dibblee (1953) suggests a right-lateral displacement of 175 miles (320 km) along the San Andreas fault in California since early Miocene time, and 350 miles (640 km) since the Jurassic.

This note presents evidence for displacements of comparable magnitude along some of the transcurrent faults of the northeastern Pacific Ocean.

The geomagnetic surveys first conducted by the Scripps Institution of Oceanography aboard the Coast and Geodetic Survey ship PIONEER and later extended by cruises of Scripps ships have revealed in the eastern Pacific Ocean magnetic anomalies with north–south lineation which are cut by the transcurrent east–west striking faults described by Menard (1959).

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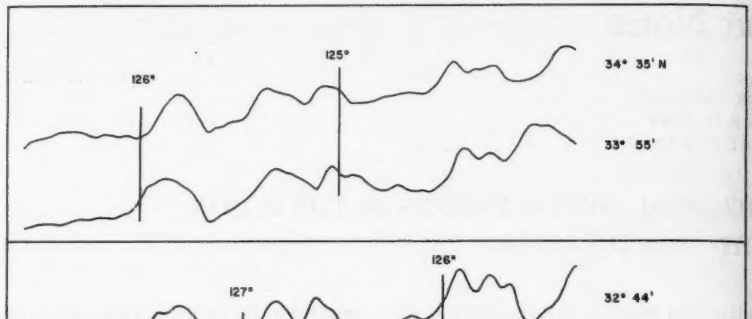


Figure 2. Profiles of anomalous total magnetic intensity north and south of the Murray fault suggesting a right-lateral slip of 84 nautical miles (154 km)

The surveyed area is shown on Figure 1. By matching the magnetic grain north and south of a fault the direction and magnitude of horizontal slip can be measured. This was first done across the Murray fracture zone (Menard and Vacquier, 1958; Mason, 1958) where a right-lateral displacement of 84 nautical miles (154 km) was interpreted (Fig. 2). The displacement along the Murray fault zone suggested an investigation of a possible displacement across the Pioneer fault 300 nautical miles to the north. The cruise of the R/V HORIZON in December 1958, undertaken for the express purpose of looking for this displacement, found a left-lateral slip of roughly 3° in longitude (Vacquier, 1959). The cruises of the R/V SPENCER F. BAIRD in August 1959 and April 1960 extended the magnetic survey to the west on both sides of the Pioneer and the Mendocino faults.

The fluxgate ASQ-3A magnetometer was used on all cruises except the last, which employed a proton free-precession magnetometer. The original data are plotted as a magnetic contour chart on Plate 1. Its eastern edge joins the western edge of the contour chart of Mason and Raff (1961, Pl. 1). The normal geographic change of the total magnetic intensity was estimated by smoothing Magnetic Chart No. 1703 published by the Hydrographic Office of the U. S. Navy. Ground-wave Loran positions taken every half hour supplemented by astronomic sights were used for locating the ship. West of 134° W. long., the reception of the ground wave becomes uncertain at night because of the interference of the sky wave. The dead-reckoned course during the night was corrected in the morning as soon as Loran signals became identifiable again. West of 140° W. long. astronomic sights and dead reckoning were the only means of establishing the ship's position. From the spread of the fixes parallel to the track, it is estimated that the average departure of the fixes from the true positions was between 1 and 2 miles. Since short-time fluctuations of the geomagnetic field were not removed from the data, fluctuations of 0.5 milligauss, the contour interval, can originate from this cause during daylight hours. Throughout the two expeditions the ship's average speed was 11 knots.

On Figure 3 profiles of the total magnetic intensity are plotted along the long east-west tracks. The profiles are drawn in their actual positions on the map. Their eastern portion extends across the map of Plate 1 of Mason and Raff (1961). The locations of the Pioneer and Mendocino faults are indicated on Figure 3. The magnetic pattern in the fault zones is too complex to permit contouring with the spac

data. In fact, the contouring of the western portion of the map of Plate 1 depends on the assumption that the magnetic trends are directed north–south as they are on the eastern half of the map where the control is adequate.

East–west magnetic intensity profiles will now be examined for fit of the magnetic grain across the faults without speculation as to origin of the anomalies. The chart of Plate 1 and the profiles of Figure 3 show that as a rule the anomalies strike N.–S. ± 10°. Occasionally a northwest–southeast trend is observed as well as circular anomalies in the fault zones. The correlation for the measurement of slip depends on the continuity of the north–south trending anomalies. Because the elongate anomalies occasionally terminate where there is no major fault, correlations across the faults cannot be perfect everywhere.

Figure 4 represents our interpretation of the magnetic data as slips along the Pioneer and Mendocino faults. The profiles north of Lat. 38°45′ N. have been translated eastward 3°05′, which amounts to 145 nautical miles (265 km). The profiles north of Lat. 40° N. have been translated eastward by an additional 13°42′ (640 nautical miles or 1160 km). The total displacement across both faults is thus 780 nautical miles (1425 km). The correlation of some of the magnetic features across the faults is shown on Figure 4 by dashed lines. To make the goodness of fit more obvious, the northernmost profile of Figure 4 has been drawn again as a dotted line between the profiles south of the Pioneer fault.

Possible origins for the observed magnetic-anomaly pattern are discussed by Mason (1958) and by Mason and Raff (1961). There is of course an infinity of possible distributions of magnetic rocks that will yield the same magnetic anomaly. However, certain types of distributions can be excluded because they would require the presence of rocks possessing unreasonable magnetic properties. To have a reasonable magnetic-polarization contrast, the bodies of rock causing the north-south trending anomalies should be at least 2 km thick. Some of these anomalies are sufficiently sharp to place the top of the magnetic rock no deeper than 6 km from sea level (1.5 km below the sea floor). A likely origin of the magnetic pattern is the intrusion of more strongly magnetic material into the basaltic crust. That the large magnetic anomalies are not directly related to the known topography is demonstrated by the anomaly caused by the 200-fathom relief of the Pioneer Ridge which is no more than one-tenth of the anomalies due to intrusives or extrusives in this fault zone, some of which are as large as 8 milligauss. Plate 1 shows a crossing of the Pioneer Ridge centering at 38°40′N. lat., 130°25′W. long., where no large magnetic anomaly was encountered.

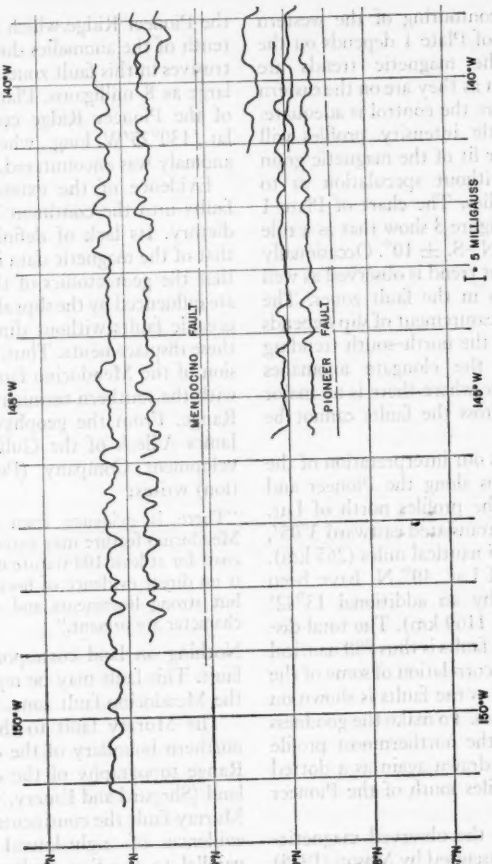
Evidence of the extension of the oceanic faults into the continent is weak and contradictory. Its lack of definiteness compared to that of the magnetic data in the ocean suggests that the geotectonics of the continental crust are influenced by the slips along the transcurrent oceanic faults without directly taking part in these displacements. Thus, the geologic expression of the Mendocino fault is that it lines up with the southern termination of the Cascades Range. From the geophysical point of view, James Affleck of the Gulf Research and Development Company (Personal communication) writes:

"There is evidence from magnetics that the Mendocino feature may extend eastward from the coast for at least 100 statute miles (160 km). There is no direct evidence of horizontal displacements, but strong lineaments and changes of magnetic character are present."

Nothing on land corresponds to the Pioneer fault. This fault may be regarded as a sliver of the Mendocino fault zone.

The Murray fault to the south marks the northern boundary of the drowned Basin and Range topography of the continental borderland (Shepard and Emery, 1941). South of the Murray fault the continental borderland shows evidence of right-lateral strike-slip faults parallel to the San Andreas fault, suggesting that the latter belongs to a fault zone a part of which is under water. On the continent the Murray fault lines up with the Channel Islands and the Transverse Ranges of California, where all horizontal displacements measured by geological methods are left lateral. This direction is opposite to the 154-km right-lateral slip found in the deep ocean by the magnetic survey. However, all magnetic and bathymetric evidence of the Murray fault stops east of 124°W. long., making a gap of 4° between the eastern end of this fault at sea and the Channel Islands.

The presence of this gap makes it possible to reconcile the opposite directions of displacement by assuming an absence of genetic relationship between the displacements at sea and on land. This explanation may be unlikely, but it follows directly from the assumption,


140°W
145°W
150°W
42°N
41°N
40°N
39°N
38°N
37°N
MENDOCINO FAULT
PIONEER FAULT
5 MILLIGAUSS

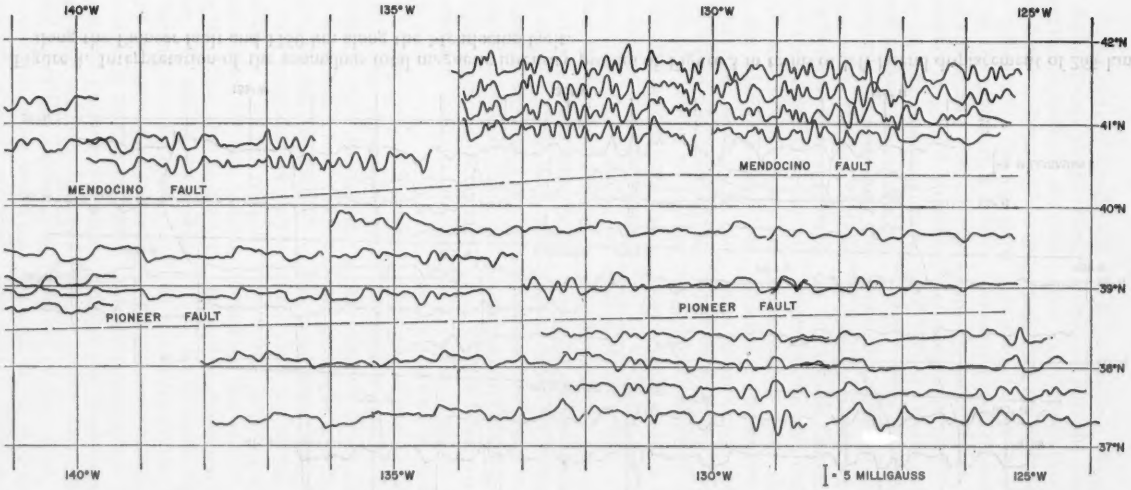


Figure 3. Profiles of anomalous total magnetic intensity along the long east-west tracks. The location of the profiles is approximately that of the corresponding tracks of Plate 1

Figure 4. Interpretation of the anomalous total magnetic intensity profiles of Figure 3 in terms of left-lateral displacement of 265 km along the Pioneer fault and 1150 km along the Mendocino fault.

which again may be unlikely although still possible, that the displacements along oceanic faults propagate into the continent at the time they happen. If this time is ancient, both magnetic and geologic evidence for the slips gets largely wiped out on the continent. Furthermore, geologists do not usually look for slips of 1400 km. The absence of definite extension of the oceanic faults into the continent can be explained by the motion along the San Andreas fault zone which separates the ocean from the continent. The land east of the San Andreas fault is now slipping to the south at the rate of 5 cm per year. Since the Jurassic it has moved 560 km (Hill and Dibblee, 1953), roughly the distance between the Mendocino and the Murray faults. There is no need to be disturbed about the discrepancy between the displacements of the Murray fault at sea and the faults on land at about the same latitude unless one insists that the oceanic faults are younger than the San Andreas fault. This is contrary to present-day evidence; the San Andreas fault is seismically active, whereas the oceanic faults are seismically dead west of 130° W. long. Also, mild heat-flow anomalies seem to cross the Mendocino escarpment as though it did not exist (R. P. Von Herzen, personal communication), indicating quiescent conditions in the last 100,000 years at least.

Recalling again that the present discussion has postulated without proof the extension of the oceanic displacements into the continent at the time of their formation, a limit to their age is set by the age of the San Andreas fault—namely, they must have stopped moving some time in the Cretaceous. At 1 cm per year, the average rate of motion along the San Andreas fault, the slip along the Mendocino fault must have taken another 100 million years. The rocks that give us the magnetic lineations that permit us to measure the lateral displacements along the transcurrent faults are thus still older and must have been laid down in Paleozoic time or earlier. They have not been disturbed appreciably since that time either mechanically or by intrusions or extrusions, or we would not have been able to read the magnetic record of the motions. Thus, except for the motions along the transcurrent faults, it is likely that the floor of the eastern Pacific Ocean between 30° and 45°N. lat. has been relatively quiescent since some time in the Paleozoic. This interpretation could be directly confirmed by dating paleontologically and geochemically the material from a deep boring in the ocean floor.

If the motions along oceanic faults do not propagate into the continent at the time of their occurrence, then nothing can be said about their age, and the preceding discussion is discarded. It can be imagined that somehow the oceanic crust carrying the magnetic pattern can slip under the continental mass without disturbing appreciably the continent's geology and isostasy. The oceanic crust would be depressed under the continent, which would cause smoothing out of the geomagnetic field as noted by Heezen (Heezen and Tharp, 1959) in the Atlantic Ocean as one proceeds shoreward over the continental slope and shelf. The same happens along the west coast of North America as shown by unpublished observations by workers at the Scripps Institution of Oceanography, except that the band of smooth field is much narrower along the Pacific margin, as one would expect it to be.

The discovery of slips of the order of 100 km in the ocean floor suggests looking for such displacements on the continents by reviewing geological and geophysical evidence. As a start, James Affleck (Personal communication) has examined for that purpose over 1 million square miles of aeromagnetic surveys. He found that there are numerous magnetic trends related to major tectonic features. Some of these can be followed for several hundred miles. In addition, offsets in anomaly-trend systems commonly suggest horizontal displacements but generally only of the order of 10 miles. In general, geomagnetic measurements indicate that the crust is more complex in continental areas than under the oceans. In evaluating the evidence from land magnetic surveys one should remember that surveys for oil exploration are confined to sedimentary basins.

The displacements along strike-slip faults in the ocean floor are thus of the same magnitude as the distances through which continents have been presumed to drift to accommodate physiographic and paleomagnetic data. Except for Carey (1958), who invokes a several-fold expansion of the whole Earth, continents are pictured in the literature as rigid plates drifting without distortion through the viscous ocean floor. Now we have evidence for the rigidity of the upper part of the oceanic crust that carries the magnetic pattern, which is just as good as the evidence for the rigidity of continents exemplified by the fit obtained by Carey (1958) between the coasts of Africa and South America along the 2000-m isobath. Since it is most unlikely that the magnetic or

the physiographic correlations are accidental, the resolution of this paradox will be an important advance in geophysics.

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MAGNETIC SURVEY OFF THE WEST COAST OF NORTH AMERICA, 32° N. LATITUDE TO 42° N. LATITUDE

Abstract: A ship-towed magnetometer survey of an area 250–300 miles wide off the foot of the continental slope along the coast of California has revealed a narrow pattern of anomalies of about 400 gammas magnitude trending north-south for more than 500 miles. The pattern is interrupted at the known faults and elsewhere; offset of the pattern at the Murray fault suggests a right-lateral displacement of about 84 nautical miles. The anomalies are such as might be expected of slablike structures underlying the ocean floor; geological possibilities include basic lava flows, topography of the main crustal layer, and intrusion of ultrabasic material from the mantle.

While making detailed hydrographic surveys off the west coast of North America, the Coast and Geodetic Survey ship PIONEER towed a total-field magnetometer and enabled us to survey an area 250–300 miles in width off the foot of the continental slope between latitudes 32° and 52° N., to a standard approaching that of air-borne magnetometer surveys over land.

A map of the section between latitudes 32° and 36° N., which includes part of the Murray fault, has already been published (Mason, 1958). This note presents a map of the area from latitude 32° N. to latitude 42° N., a little north of the Mendocino fault, and includes the section published previously. The geographical location and major physiographic features of the area are shown in Figure 1.

The area surveyed was covered with a regular grid of lines, normally about 5 miles apart (about twice the average depth of the water). Except in the vicinity of the Mendocino fault, where intersecting lines were run north–south and east–west, the lines were mostly east–west —*i.e.*, across the grain of the anomalous magnetic field. Sufficient lines were run in the other direction, however, to resolve any ambiguities that might otherwise have arisen in the contouring process. A radio-navigation system, using fixed beacons ashore, enabled positions to be determined about every 8 miles with a probable error of the order of 0.1 mile.

The magnetometer was a highly stabilized, self-orienting, total-field fluxgate. To reduce the effect of the ship's magnetism, the fluxgate mechanism was mounted in a streamlined fish and towed 500 feet behind the ship. The instrument was calibrated ashore, at approximately monthly intervals, by comparison with a proton-precession magnetometer. Taking into account the principal sources of error, namely inherent drift, inaccuracy of adjustment, and the effect of the ship, the standard deviation of a single observation from the absolute value is about 15 gammas. Provided that the instrument is kept in proper adjustment, the relative error between any two points on the same line is negligible.

The magnetic anomaly map (Pl. 1) is a plot of the observed field from which the regional magnetic field has been subtracted. The latter was derived from the map of the observed field by a smoothing process. No attempt was made to remove the effects of diurnal variation and magnetic storms. The daily range in hourly values of the total field at these latitudes is usually of the order of 30 gammas, and seldom exceeds 50 gammas, except during magnetic storms. It is unlikely, therefore, that the normal diurnal variation will have any significant effect on the positions and magnitudes of the more important anomalies, although individual contours may be distorted, particularly in areas of shallow magnetic relief. There were several magnetic storms during the survey, but the contours did not appear to be significantly affected by them. Long-term changes in the earth's magnetic field are automatically removed in the smoothing process for deriving the regional field.

The regional bathymetry has been described by Menard and Dietz (1952) and by Menard (1955). The area is crossed by two of the great east–west fault zones of the northeast Pacific,

Geological Society of America Bulletin, v. 72, p. 1259–1266, 4 figs., 1 pl., August 1961

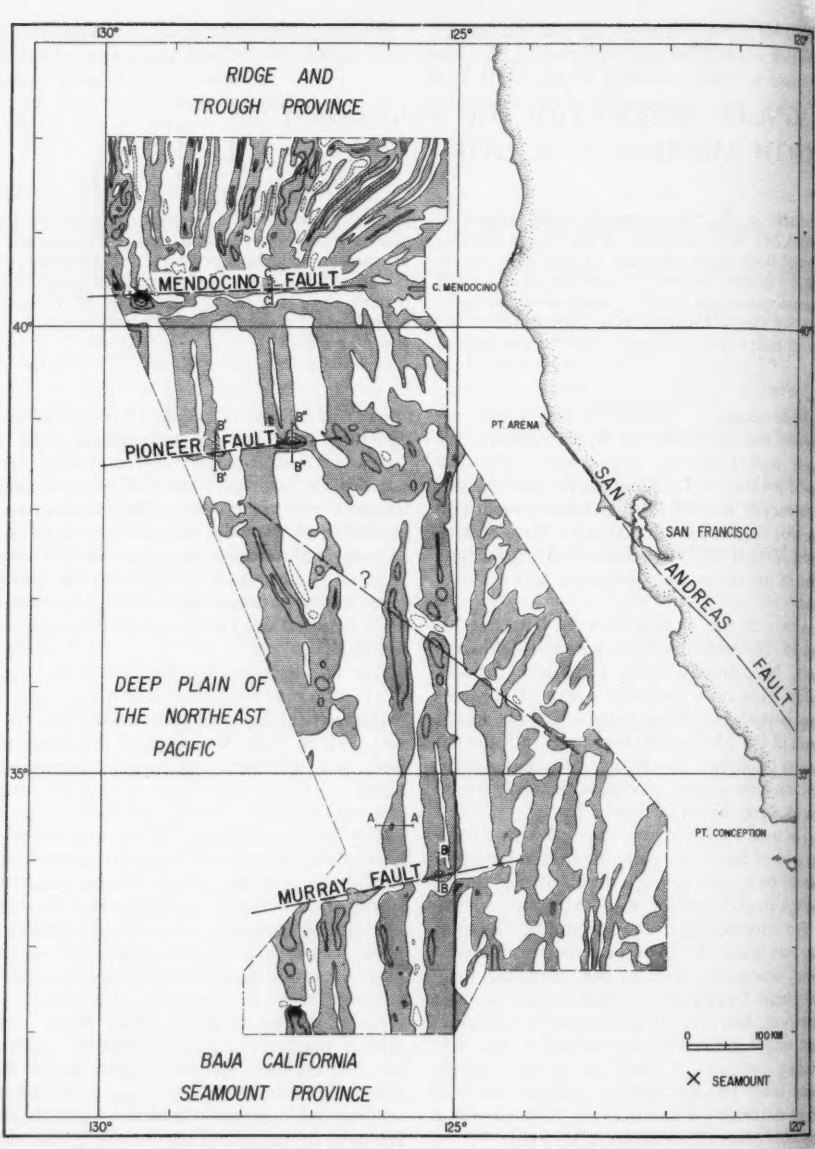

Figure 1. Skeleton magnetic map, showing geomorphic provinces, principal known faults, and locations of profiles. The contour interval is 250 gammas. Areas of positive anomaly are stippled. Negative contour lines are broken.

the Mendocino and the Murray. The former is by far the most spectacular topographic feature of the area; the bottom rises about 800 fathoms (1½ km) from south to north across it, and across most of the map it forms an asymmetrical ridge with a south-facing escarpment 5000–10,000 feet (1½–3 km) high. It marks the boundary between the Deep Plain of the Northeast Pacific, a comparatively smooth abyssal plain 2100–2600 fathoms (4-5 km) deep, to the south, and the shallower Ridge and Trough province, an area of narrow north or northeast trending ridges and troughs, to the north. The Murray fault, about 6° south of the Mendocino, divides the Deep Plain from the slightly shallower Baja California seamount province. A recently discovered topographic feature, the Pioneer Ridge, crosses the Deep Plain about 100 nautical miles (185 km) south of the Mendocino fault, marking the line of another major east–west fault.

Seismic-refraction measurements have been made at points widely distributed throughout the area. Results for stations south of the Mendocino fault are consistent with a normal oceanic crust (R. W. Raitt, personal communication). Thus at A–A (Fig. 1) the structure consists of 4.6 km of water, 0.3 ± 0.1 km of unconsolidated sediments, 1.4 ± 0.3 km of a second layer of velocity 5.4 ± 0.4 km/sec and commonly believed to be either volcanic rocks or consolidated sediments, and the main crustal layer 4.8 ± 0.5 km thick and of velocity 6.9 ± 0.1 km/sec. The M discontinuity is at a depth of 11.1 ± 0.6 km. North of the Mendocino fault, where the bottom is shallower, the structure appears to be distinctly different, but the exact nature of the difference has not yet been established.

The magnetic map is unusual in many respects. Although some correlation might have been expected between the magnetic field and the linear topography of the Ridge and Trough province, there were no grounds for anticipating either the strongly lineated character of the map over the Baja California seamount province and the relatively featureless Deep Plain, or the large horizontal displacements of the magnetic pattern that occur at the known faults and elsewhere. These lineations, and the horizontal displacements, far exceed in scale any that have been observed over the continents.

At the Murray fault, a 200-km section of the pattern on the north side can be matched against a similar section on the south side in a way that suggests a right-lateral displacement of about 84 nautical miles (155 km) (Mason, 1958). There is no obvious match of the pattern across either the Pioneer or the Mendocino fault, within the limits of the map. The survey has recently been extended westward on both sides of these faults, by ships of the Scripps Institution of Oceanography, to look for such a match and hence to determine the strike-slip movement along them. The results are reported elsewhere (Vacquier, 1959; Vacquier *et al.*, 1961). They are interpreted as indicating a left-lateral slip of about 140 nautical miles (265 km) across the Pioneer fault and about 640 nautical miles (1185 km) across the Mendocino.

At the Mendocino fault the character of the map changes abruptly. Some of the anomalies on the north side are so sharp as to place the top of the magnetic rocks no deeper than 3½ km (1900 fms) below sea level, which is shallower than the sea floor on the south side of the fault, and shallower therefore than the magnetic structures on that side. The change in character of the map is undoubtedly due in part to this factor. However, the seismic results, the change in elevation, and the fact that the line of the Mendocino fault marks the approximate boundary between various physiographic provinces of the neighboring continent (Eardley, 1951, p. 430) all point to a more general change in crustal structure. There is no significant change in the character of the map across either the Pioneer or the Murray fault.

The lineations of the anomalous magnetic field are remarkably straight. The principal north–south trending anomalies, for example, strike almost exactly N.5° W. from the southern edge of the map to the Mendocino fault, a distance of 500 nautical miles (900 km). This is surprising, in view of the very large relative displacements that have taken place at the intervening faults. They are also remarkably uniform in magnitude and width. Over most of the area they show little correlation with topography, and there is no indication of any systematic difference between seismic velocities in the anomalous areas and in the areas between, although the amount of data is not yet sufficient to demonstrate this point conclusively.

As to their immediate cause, the anomalies are such as might be expected of flat slablike structures approximately underlying the areas of positive anomaly. Figure 2 shows the ob-

served anomalies along lines A–A and B–B of Figure 1, and superimposed on them theoretical anomalies computed respectively for an east–west line across an infinite north–south striking slab, and for a north–south line across the southern edge of a semi-infinite slab, both immediately underlying the ocean floor. The agreement is striking.

Two factors restrict the theoretically infinite range of possible depths and thicknesses of these slabs: (1) the sharpness of the anomalies places a limit on the depth to the magnetic rocks; and (2) since there is an upper limit to the polarization contrast that can reasonably be assumed, the bodies must be of a certain minimum thickness to produce anomalies of the observed magnitude.

The effect of these restrictions was investigated by comparing profiles from various parts of the map with theoretical anomalies computed for slabs of various depths and thicknesses. This investigation showed that the anomalies can be closely matched by slabs lying in the "second layer," and by slabs extending through the whole depth of the crust but that, in general, anomalies computed for

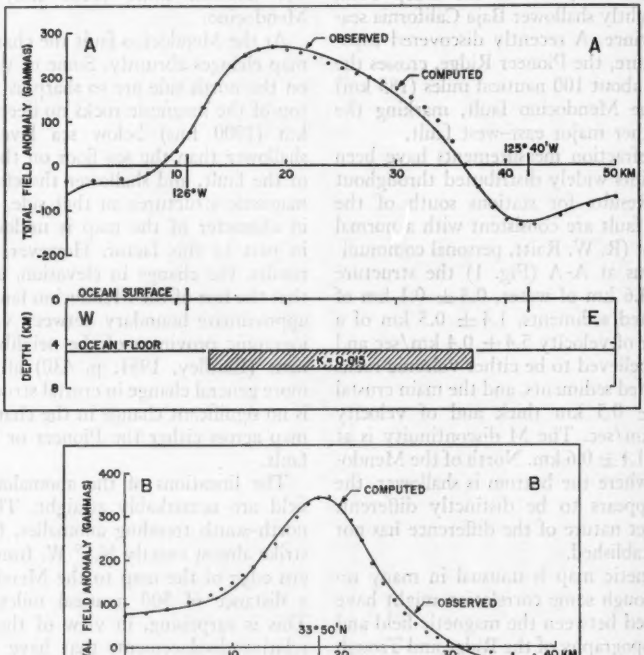

Figure 2. Observed anomalies along lines A–A and B–B of Figure 1. Superimposed on them are theoretical anomalies computed respectively for an east-west line across an infinite north-south striking slab, and north-south line across the southern edge of a semi-infinite slab.

slabs confined to the main crustal layer do not match the sharpness of the observed anomalies. Polarization contrasts corresponding to susceptibilities in the range 0.005–0.015 are necessary, which are values appropriate to the contrast between highly magnetic basalts and relatively nonmagnetic sediments.

both the second and the main crustal layers. Objection might be raised to all three on the grounds of lack of topographic and seismic expression. Figure 3 illustrates the three cases in terms of infinite north–south structures each of which has been adjusted by trial and error to fit the anomaly exactly (Mason, 1958).

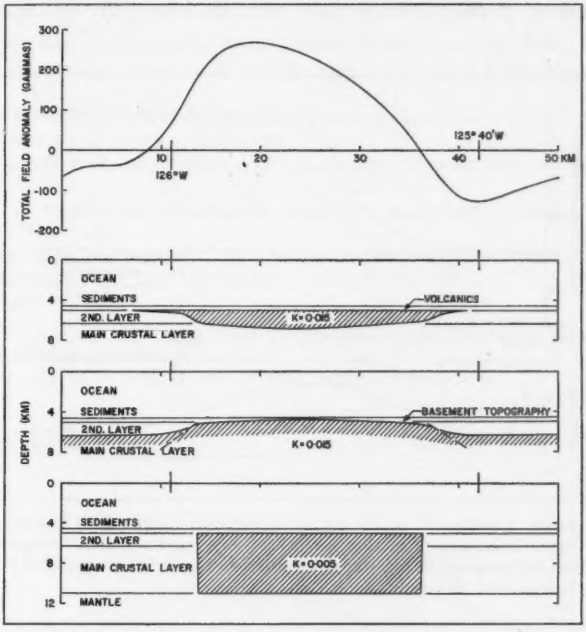

Figure 3. Possible interpretations of magnetic profile A–A. The theoretical anomaly computed for each of the three infinite north-south striking structures fits the observed anomaly almost exactly.

The geological possibilities fall into three categories: (1) isolated bodies of magnetically anomalous material, for example basic lava flows, within the "second layer"; (2) elevated folds or fault blocks of the main crustal layer; and (3) zones of intrusion of highly magnetic material from the mantle, extending from top to bottom of the crust. Category (1) would fit in well with current views about the somewhat enigmatic "second layer." Such lavas would constitute 10 per cent of the local crustal volume. For (2), a parallel might be drawn with the fault-block topography of the neighboring Basin and Range province. For (3), the anomalies might arise from magnetic contrasts within

On the magnetic map the Mendocino fault is marked by a narrow east–west anomaly of 300–500 gammas, almost symmetrical about the crest of the ridge. Figure 4 shows a typical section across the fault, along line C–C of Figure 1, and the observed anomaly along the same line. The question arises as to whether the anomaly is directly related to the ridge—*i.e.*, whether the magnetic rocks conform to the surface topography. Figure 4 shows a theoretical anomaly computed for uniform magnetization of the topographic section, assumed infinite in the east–west direction. The discrepancy between observed and computed anomalies suggests a deficiency of magnetic

material on the south side of the fault. The stepped section, which has been adjusted by trial and error to approximately fit the anomaly, and which agrees closely with a possible interpretation of a nearby gravity profile (J. C. Harrison, personal communication), suggests a steeper and more abrupt discontinuity than might be expected from the topographic profile. For a polarization contrast of 0.007, a vertical displacement of about 3 km is indicated, or proportionately more or less for smaller or larger polarization contrasts.

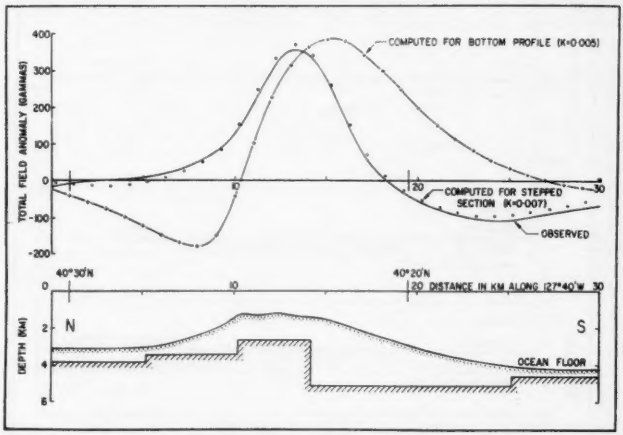

Figure 4. North-south profile across the Mendocino fault along line C–C of Figure 1, showing theoretical anomalies computed for uniform magnetization of topographic and stepped sections respectively, assumed infinite in the east-west direction

The vertical relief across the Pioneer and Murray faults averages only a few hundred fathoms; the Pioneer fault forms a shallow ridge with a south-facing scarp, the Murray rises irregularly from north to south. On the magnetic map these faults are marked more by the effects of the transcurrent movements than by any persistent feature that could be ascribed to vertical displacements. The elliptical pattern of contours at B–B of Figure 1 on the line of the Murray fault, and the similar patterns at B′–B′ and B″–B″ on the line of the Pioneer fault, can be interpreted as the effects of south-facing ends of blocks intersected by the faults. The negative anomaly under the "A" of Murray in Figure 1 indicates a north-facing end.

This discussion has necessarily been based on very simple models. Not only are individual structures likely to be complex, but considerable heterogeneity of magnetic polarization is to be expected also. However, from their overall uniformity, the anomalous trends must clearly be considered part of a single homogeneous pattern. In the absence of any magnetic structures cutting across them, except near the foot of the continental slope, where there is evidence for complex faulting with north–south components of slip, it must be concluded that the faults are more recent than the structures. Possibly the latter are relatively ancient features of the crust. A more significant discussion of both pattern and faults will be possible when the rest of the map, and projected extensions westward and over the continental shelf, have been completed, and when samples of the rocks have been obtained.

Acknowledgments

The authors gratefully acknowledge the cooperation of the Coast and Geodetic Survey and particularly wish to place on record their appreciation of the interest taken in the work by Captain (now Admiral) Charles Pierce, who was in command of the PIONEER during the early stages of the survey. They also wish to

acknowledge the facilities for calibrating the magnetometer generously made available by the Varian Associates of Palo Alto, California, and the interest taken in the work by Drs. L. L. Nettleton and N. C. Steenland of the Gravity Meter Exploration Company of Houston, Texas. They are indebted to colleagues at Scripps for helpful discussion of the results, and to Dr. R. W. Raitt, in particular, for supplying the seismic data. Dr. J. C. Harrison of the Institute of Geophysics supplied unpublished gravity data. The work was supported by the U.S. Office of Naval Research under contract Nonr–233(04).

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MAGNETIC SURVEY OFF THE WEST COAST OF NORTH AMERICA, 40° N. LATITUDE TO 52° N. LATITUDE

Abstract: An extensive high-resolution magnetic survey of total field at sea level reveals unusual north-south lineations and much crustal faulting. Computations indicate that the linear pattern is due to strongly magnetized mafic rocks beneath the sediments.

A magnetic survey was made off the west coast of the United States and Canada, 40° N. lat. to 52° N. lat. as a continuation of work reported by Mason (1958) and Mason and Raff (1961). The total field at sea level was measured by a fluxgate magnetometer towed by a ship that ran east–west track lines spaced 5 nautical miles apart except for highly anomalous areas where the spacing was about 1 mile. The accuracy of navigation was 0.1 mile. The area surveyed is mostly over deep water (1500–1700 fathoms) and covers 230,000 square miles. Since the diurnal variation is small compared to the value of the anomalies, its effect was not removed. The map (Pl. 1) was contoured with intervals of 100 gammas, whereas the map by Mason and Raff (1961) has an interval of 50 gammas.

The unique north–south trends of the magnetic contours of Plate 1 make it possible to see many faults in the ocean's crust even though this area is covered with deep continental sediment (Menard, 1960). In locating and measuring these faults we assumed that the north–south trend was originally continuous and predated the faults. The faults are observed as a line of anomalous features, as a discontinuity of contour density, or as a shear of the north–south lineations.

At about 40° N. lat. (Fig. 1; Pl. 1) the strong east–west feature is the Mendocino fault, which Vacquier and others (1961), by extending the survey of this fault to the west, have shown to have a left-lateral slip of approximately 600 nautical miles. Two subsequent magnetometer lines run between the eastern limit of the present survey and the coast show that the Mendocino anomaly continues to the coast. This great fault is here a boundary between provinces of different ocean depth and different magnetic anomaly density. North of the Mendocino fault (41° N. lat.) and near the eastern limit of the survey, the lineations turn to the northeast as though the crust had been rotated by the action of the several large faults that converge in this area. Along the line A–A (Fig. 1) the linear structures bend smoothly, which is unusual. The anomalous interruption just northwest of line A–A has associated with it a small basin 400 fathoms deeper than the surrounding area.

The anomalous feature along line B-B has no topographic counterpart. The angular and curved line C-C-C marks a boundary of gradient density similar to that usually found along the continental shelf, but the continental shelf does not show any such departure here. Line D-D marks a row of three seamounts. Lines E-E, G-G, and H-H mark a boundary of magnetic gradient-density change that is not indicated by the topography. The area between these lines appears from the magentics to be either smoothed or depressed with respect to adjacent aras. The area enclosed by the B, E, G, and H lines looks like a dilated section with the motion hinged at the western corner. Lines J-J and K-K mark a fault of the magnetic linear trend. These two lines may be parts of the same fault that has disturbed a wide area. There is a suggestion from the map that the magnetic lineations L-L, M-M, and N-N are all the same feature with the center section having slipped to the east. Most of the area on the map from 41° N. lat. to 49° N. lat. looks like angular pieces of crust that have slipped and rotated with respect to neighboring pieces. The anomaly along Q-Q is associated with a trough that is 800 fathoms deeper than the surrounding area. Along the line R-R there is a 500-fathom ridge that has little or no magnetic counterpart.

Along the line S-S-S there is a suggestion of a fault, but nothing is indicated from the bottom

Geological Society of America Bulletin, v. 72, p. 1267–1270, 1 fig., 1 pl., August 1961

topography. In the general region west of this line there are many seamounts, some of which cause strong magnetic anomalies, while others do not. It is not at all clear why volcanic seamounts in the same area should give magnetic anomalies of such different magnitudes. Farther north the irregular line T-T encloses an area of apparent magnetic depression similar to the one to the south. There are a number of other partial breaks in the linear pattern such as along lines U-U, V-V, and W-W.

By far the largest topographic feature in the area of the survey is the continental shelf, and it shows little magnetic anomaly, as would be expected from its shape in cross section at the places where the magnetometer has crossed it. It appears to give a slight negative anomaly, suggesting that the material of the oceanic

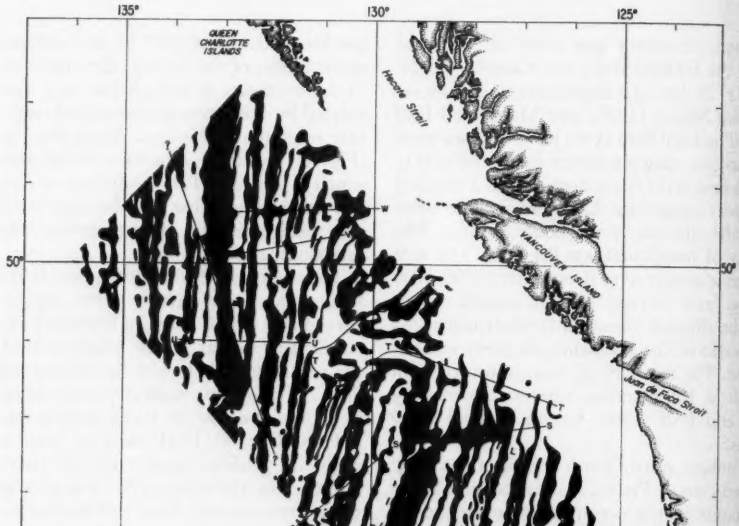

Figure 1. Index anomaly map of the total magnetic field. The positive area of the anomalies is shown in black.

crust is much more magnetic than that of the continental crust.

These north-south linear features have aroused considerable interest and speculation. They disappear over the areas where the continental crust extends into the ocean. The layer containing these magnetic structures may dip under the continental crust much as the M discontinuity does; this is supported by the changes in slope width and gradient of the magnetic ridges where they occur closer to the shore line. Using rules of thumb to find the depth of burial of the structures the structures are placed about 3 times deeper at the continental shelf than they are farther out to sea. The apparent deepening of these structures near the coast could be an illusion, however, caused by their having a wider thinning edge near the shelf.

There is as yet no satisfactory explanation of what sort of material bodies or physical configurations exist to give the very long magnetic anomalies. It has been suggested that they are troughs and ridges buried under sediment, platelike lava flows that filled older troughs, unique groupings of vertical dikes, high iron-content derivatives of weathering that filled older troughs, or a thermal pattern that has "highs" and "lows" about the Curie point of ocean rocks. Two of the principal difficulties in imagining a body that could cause these north-south features are that the measured magnetic anomalies and computations about them require that the upper surface of the body be within 1 km of the sedimentary layer and that the body have what is considered an unusually strong magentic moment. Such features were not revealed on the basis of topography nor have they been detected by seismic methods (R. W. Raitt, personal communication).

Magnetic measurements now being made of all the rocks collected in the Scripps dredge hauls indicate that most oceanic igneous rocks have a remanent magnetization and susceptibility several times greater than the median value for continental igneous rocks. Thus, a predominance of oceanic igneous rocks are similar magnetically to the less common mafic rocks of the continents. Magnetic patterns over the oceans should then show greater magnetic anomaly than generally is observed on land, and possibly, although masses of rock may be too thin to be detected seismically, they may have sufficient magnetic moment to give a strong anomaly.

If there originally had been a trough and ridge system, these troughs and ridges should show at least slightly in the topography of the heavily sedimented ocean floor. If the magnetic bodies are the original ridges, the magnetic "highs" should correspond with the slight topographic ridges; but if the magnetic bodies are accumulated material in the valleys, then the magnetic "highs" should be over the slight topographic valleys. A statistical correlation analysis made of the magnetic and topographic relief for some of this area showed that the topographic ridges and magnetic "highs" correspond in about 65 per cent of the cases as against 35 per cent of negative correspondence. This is certainly not conclusive, but does suggest that the magnetic bodies are the original igneous ridges or elevated areas that survive in a modified, perhaps flattened, form.

As to the notion that this is a thermal pattern about the Curie point, heat-flow measurements (Von Herzen, 1959) give values that make it almost impossible for such temperatures and thermal gradients to exist close to the sedimentary layer.

Acknowledgements

The success of this survey was made possible by the interest and help of many people: the officers and men of the Coast and Geodetic Survey aboard the ship PIONEER, a number of people at Scripps Institution of Oceanography who traveled with the field instruments or worked at reducing and presenting the data, and Varian Associates who helpfully furnished absolute calibration for the fluxgate magnetometer used at sea. About 5 per cent of the track data was collected by ships of the Scripps Institution of Oceanography, and another 4 per cent of the track data was obtained aboard the REHOBOTH, sailing for the U.S. Hydrographic Office.

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ARTHUR D. RAFF
E. L. JONES

EXPLORER BANK—A NEW DISCOVERY IN THE CARIBBEAN

Abstract: Scientists on the Coast and Geodetic Survey Ship EXPLORER discovered and investigated an "atoll-like" bank rising from nearly 1000 fms to 15 fms about 95 miles off the Caribbean coast of Honduras. A towed magnetometer showed that the primarily calcareous bank has a core of igneous origin.

During the 1960 oceanographic expedition of the Coast and Geodetic Survey Ship EXPLORER a previously uncharted bank, structurally comparable to the Pacific atolls, was discovered in the Caribbean Sea, and a reconnaissance survey was made. The feature lies 95 nautical miles off the coast of Honduras centered at 16° 55′ N, 83° 15′ W, and rises abruptly from general depths of nearly 1000 fathoms to a least depth of 15 fathoms (Fig. 1). It measures 5½ by 12 nautical miles within the 20-fathom contour and is essentially flat-topped except for a slight rim found on the two crossings of the eastern edge. A towed magnetometer belonging to the Scripps Institution of Oceanography revealed a core of igneous rock and indicated that the bulk of the feature is nonmagnetic. A cored rock sample from the crest was coralline and algal limestone. The bank is significant because there are few "atoll-like" structures in the Atlantic and because a magnetic survey could be made of this drowned feature.

The feature, tentatively called Explorer Bank, is an apparently isolated feature rising more than a mile above the sea floor 45 nautical miles southeast of the nearest land, the Swan

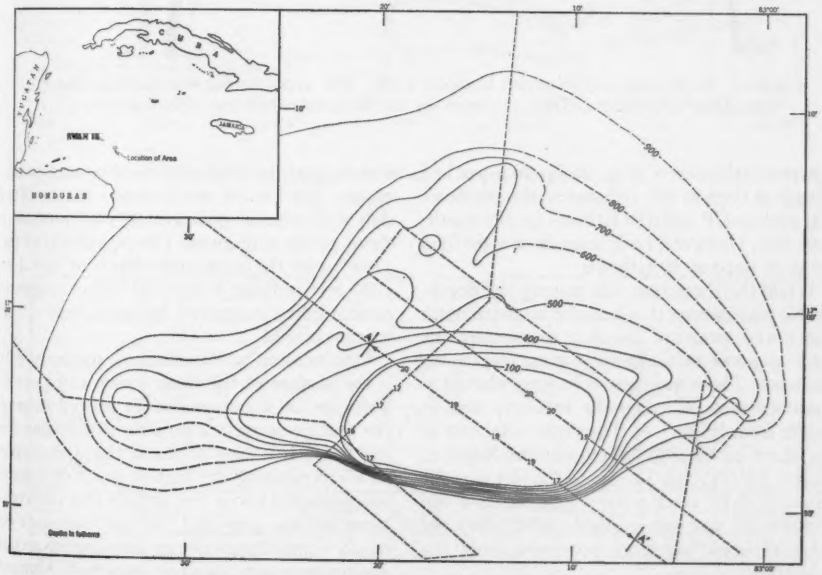

Figure 1. Reconnaissance bathymetric chart of Explorer Bank. A–A′ is the profile shown in Figure 2

Geological Society of America Bulletin, v. 72, p. 1271–1274, 4 figs., August 1961

Islands. The top of the feature is essentially rectangular, 5½ by 12 nautical miles, with the long axis oriented east-west (Fig. 1). The southern side has a measured lower slope of 31° which becomes 27° nearer the crest. To the north the slopes are gentle. The crest of the feature is nearly flat at depths of 19 to 20 fathoms. The two crossings of the eastern edge showed a narrow rim rising 2 to 5 fathoms above the general level of the flat top. The northwest-southeast profile across the bank as recorded on

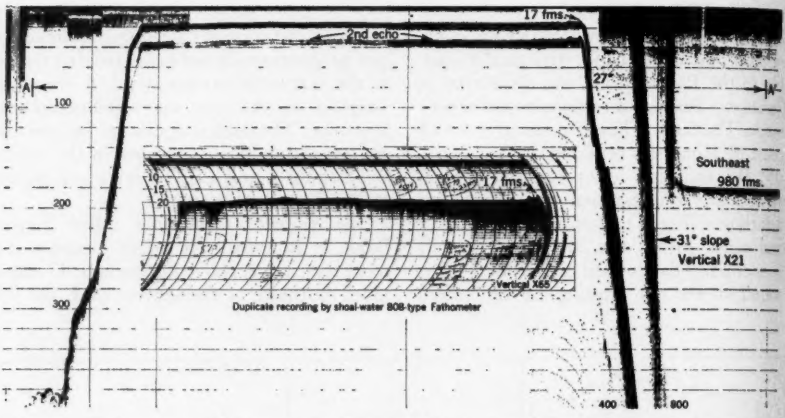

Figure 2. Fathogram profiles across Explorer Bank. The upper profile was made by the Precision Depth Recorder (PDR), the lower by the shoal-water 808-type echo sounder

the two fathometers (Fig. 2) shows a marked change in slope at 160 fathoms on the southeast side and at 270 and 310 fathoms on the northwest side. However, both sides show a decided break in slope at 40 fathoms.

While the EXPLORER was making the depth-profile measurements, a fluxgate magnetometer was towed astern of the ship to measure the total magnetic-field intensity at sea level along the track. These magnetic data were plotted as contours of equal magnetic intensity and in profile form (Figs. 3, 4). The slight difference in the shape of the 500-fathom contour between Figure 1 and Figure 3 is due to the fact that the adjusted ship's track from which Figure 1 was constructed was not available until after the magnetic data had been processed using the field track.

The magnetic-intensity contours and the profile indicate that the bulk of the feature is of nonmagnetic material, most likely limestone, as was the one sample recovered from the flat top. The sharp minimum in the magnetic profile locates a core of material of igneous origin and of shallow depth. This part of the magnetic profile is drawn as though smooth but actually is composed of small peaks and troughs. A study of these small features on the original record indicates that the highest part of the magnetic rock is less than 1200 feet deep, but probably deeper than 500 feet. The fact that the sharp feature of the magnetic profile is a "low" at first suggests rock of negative magnetic polarization, but further study shows that a tilted slab with normal polarization may produce a sharp profile minimum. The magnitude of the "low" and the maximum depth of burial of 1200 feet indicate a material of low magnetic susceptibility compared to basalt, but yet of igneous origin.

One solution to the magnetic measurements is the picture of the bank shown in Figure 4. This internal structure and the general shape of the feature suggest a possible explanation for this bank. It is most probable that the igneous rock representing the highest point of a tilted slab extended above the surface of a relatively lower sea and provided the hard substrate on which a community of reef-building calcareous marine organisms became attached. Upward growth of the calcareous material continued and kept pace with a relative rise in sea level until the igneous material was covered. Relative sea

level must have continued to rise with a concomitant upward growth of the calcareous material at least to its present extent of more than a mile high.

Possibly the flat top results from marine planation of the calcareous material that previously existed above sea level. However, the writers feel that it is more probable that this flat top represents the original surface of the living reef. In this case, the definite break in

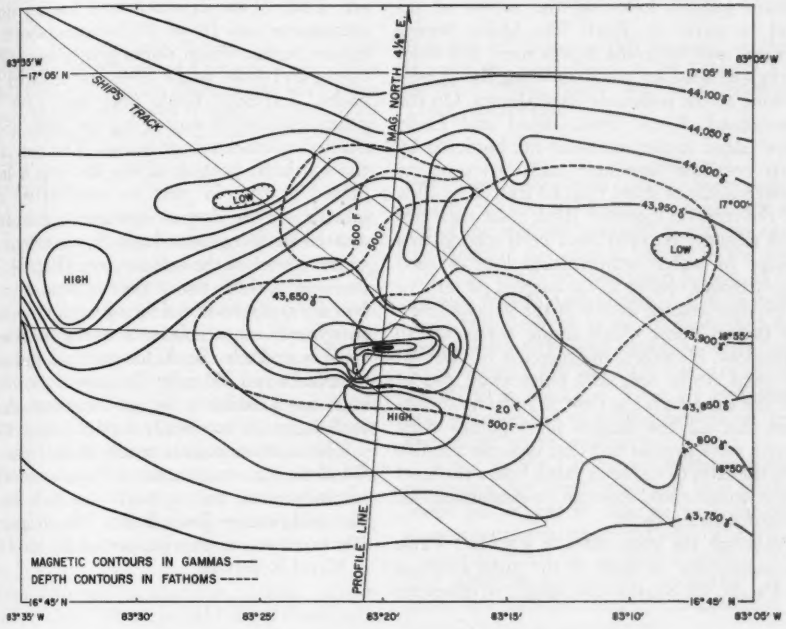

Figure 3. Magnetic field contour map of Explorer Bank

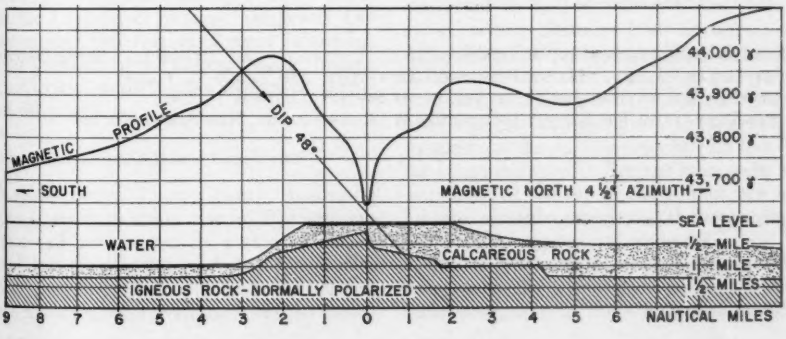

Figure 4. Profile of Explorer Bank with undistorted vertical scale

slope at 40 fathoms may be an indication of the effect of Pleistocene lowering of sea level which merely indented the sides rather than planing off the entire top of the reef. If the top of the reef was exposed above sea level, then surely it was killed by this exposure. The question then arises as to why this reef did not resume its upward growth following the return of sea level to cover it. Reefs like Quita Sueño, Serrana, and Serranilla Banka some 200 miles to the southeast are now flourishing living reefs growing at the present level of the sea. On the other hand, Great Swan Island and Little Swan Island to the northwest are bordered by steep coralline limestone cliffs attesting to relative uplift of more than 60 feet. Why then did the reefs on Explorer Bank cease their upward growth? It could have been related to a change in water temperature, loss of food supply from a change in the current pattern, or an extreme change in any of the various limiting factors upon which living corals are so dependent. However, the presence of flourishing coral banks only 200 miles away and of uplifted coral reefs less than 50 miles away suggests that all the factors for vigorous coral growth were present but that tectonic activity kept the crest of Explorer Bank below the level where corals could become re-established for successful reef growth.

Although the steep sides of Explorer Bank are comparable to those of the great atolls of the Pacific, the top does not appear to represent a true atoll. The narrow rim on the east side rises only 2 to 5 fathoms above a nearly flat top and was not found at the six other places where the edge of the reef was crossed. Alexa Bank north of the Fiji Islands in the Pacific is a true drowned atoll in which the original reef top is preserved at 20 fathoms (Fairbridge and Stewart, 1960). That drowned reef has an almost continuous rim 10 to 15 fathoms above the former lagoon which slopes gently toward the center and from whose floor rise isolated pinnacles. Explorer Bank does not have this prominent rim or sloping lagoon floor, and no isolated pinnacles were found. The rim along the southeast portion of the flat top is interpreted as due in part to accelerated coral growth on the east or upcurrent side as indicated by the greater bank development east of the center of the igneous core (Fig. 3). The rim may also be due in part to sand cays and boulder ramparts that develop on the windward margins of comparable features elsewhere in the Caribbean. Pedro Bank, for example, just south of Jamaica and 300 miles due east of the present area, has a similar string of linear and arcuate reefs along the southeastern edge (Zans, 1958).

This work represents results of the Coast and Geodetic Survey Explorer Expedition whose scientific work was supported in part by the National Science Foundation. The magnetometer work was in part supported by the Office of Naval Research.

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Contribution from The Scripps Institution of Oceanography, New Series No. 187

HARRY G. RODIS

USE OF WATER-WELL DATA IN INTERPRETING OCCURRENCE OF AQUIFERS IN NORTHEASTERN LYON COUNTY, MINNESOTA

Abstract: In northeastern Lyon County the areal distribution of aquifers of Cretaceous age determined from well-bottom altitudes suggests a series of interbedded sandstones striking northwestward and overlapping one another to the northeast. Probably the numerous thin sandstone aquifers in the area were deposited near the flanks of a Precambrian granite "high" area by a transgressing Late Cretaceous sea.

Introduction

Most domestic water supplies in northeastern Lyon County, Minnesota, are obtained from Upper Cretaceous sandstones at depths ranging from about 100 to 300 feet. This paper shows how a simple statistical evaluation of well depths suggested a possible depositional environment for these sandstone aquifers.

The strata extend from Redwood County northwestward through northeastern Lyon County into Yellow Medicine County. Records of wells outside this area show these sandstones to be scattered in occurrence and limited in extent. In Lyon County they underlie an area approximately 20 miles long and 8 miles wide and occur between altitudes of 890 and 1020 feet.

Data collected in this area include records of more than 300 wells; 212 are drilled wells completed in the Cretaceous sandstones, and the remaining are shallow bored or dug wells finished in glacial drift. The well data were collected in conjunction with a countywide study of the geology and ground-water resources by the U. S. Geological Survey in co-operation with the Division of Waters, Minnesota Department of Conservation. Robert Schneider, district geologist of the Ground Water Branch of the Geological Survey for Minnesota, supervised the work.

Geologic and Ground-Water Conditions

In northeastern Lyon County, horizontal Upper Cretaceous strata rest on Precambrian granite and are covered by a thin veneer of glacial drift (Fig. 1, section A-A′). The granite surface rises from about 750 feet to 1100 feet above mean sea level from southwest to northeast. The Upper Cretaceous strata are approximately 350 feet thick near the southwest margin of the area and gradually pinch out against the flanks of the granite "high" in the northeast. The strata are composed largely of thick sections of soft dark-bluish-gray shale but include thin beds of loosely consolidated sandstone. The interbedded sandstone strata, which occur between 890 and 1020 feet, average only a few feet in thickness. They are composed of fine, subrounded to subangular quartz grains and are of low permeability.

Water in the Cretaceous sandstones occurs under artesian conditions, and it is generally softer but higher in chloride than the water from other formations in the county (Rodis and Schneider, 1960, p. 1). Local drillers report that wells are completed "open end" in the upper few inches or feet of the sandstone and yield 2–7 gallons per minute.

Statistical Evaluation of Well-Bottom Altitudes

A simple statistical study of well-bottom altitudes was made to interpret the mode of occurrence of the sandstones. Well depths were converted to well-bottom altitudes with the aid of surface altitudes from an unpublished Army Map Service topographic map having a 25-foot contour interval. The interpolated land-surface altitude was weighted against a brief topographic description of each well site to minimize the possible error of 12.5 feet (half the contour interval). Because most wells were completed at the top of the aquifer, the well-bottom altitude represents the approximate top of the water-bearing sandstone.

The well-bottom altitudes were plotted on two graphs (Fig. 2), for areas east and west of a shale re-entrant in the Cretaceous sandstone

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which partitions the area. Each small solid circle on the graphs represents one well completed at the indicated altitude. Some errors may have been incurred in the interpolation of surface altitudes and use of reported well depths; however, because large numbers of well-bottom altitudes were plotted, the writer believes that significant results were obtained.

Well-bottom altitudes shown on Figure 2 range in altitude from 890 to 1020 feet. Between 930 and 960 and 965 and 975 on the west graph and between 930 and 962 on the east graph, the well completions appear to increase in density. These intervals probably represent individual water-bearing zones; absence of water-bearing zones is shown by intervals where there is a marked scarcity of wells. The absence of wells completed at altitudes of 965 to 975 feet on the east graph and the marked increase at that interval on the west graph in the number of well completions at the corresponding altitudes suggests that sand was being deposited in the west while clay was being deposited in the east.

The well-bottom altitudes were grouped more or less arbitrarily into three vertical intervals to facilitate plotting the geographic distribution of the wells completed at approximately the same altitudes or vertical intervals (Fig. 2): a triangle was used for wells whose well-bottom altitudes ranged from 890 to 925 feet, a cross for those between 930 and 962 feet, an open circle for those between 965 and 975 feet, and a solid circle for wells between 978 and 1020 feet (Fig. 1, map). The geographic and stratigraphic distribution of well symbols suggests a steplike structure of the sandstone in the subsurface: wells symbolized by a triangle

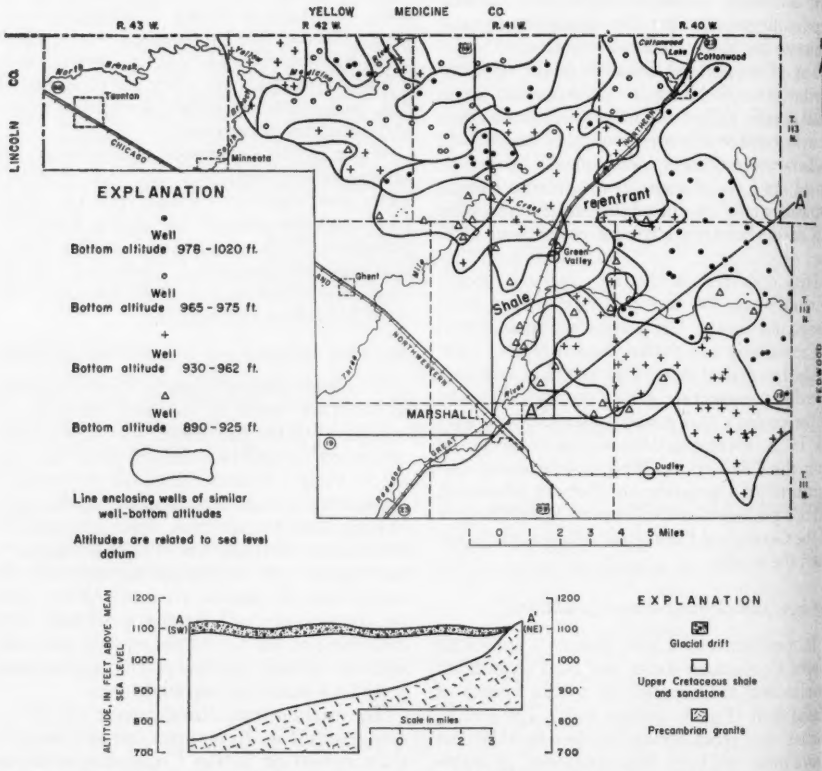

Figure 1. Map and section of northeastern Lyon County, Minnesota, showing areal distribution of well-bottom altitudes and area in which most wells were completed in Upper Cretaceous sandstone

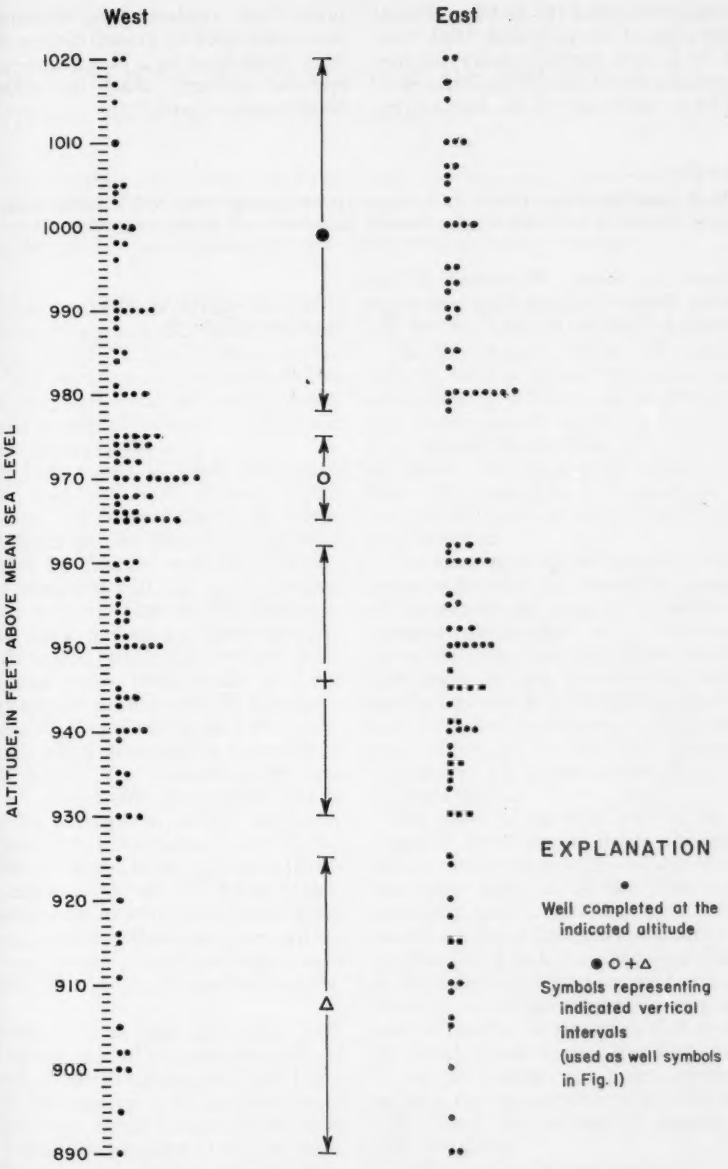

Figure 2. Arithmetic graphs showing well-bottom altitudes of wells located east and west of re-entrant that partitions report area

are concentrated around the southern rim and along the edges of the re-entrant. Wells symbolized by a cross generally occur between those symbolized by a triangle and those symbolized by an open circle in the west and between those symbolized by a triangle and those symbolized by a closed circle in the east. Wells symbolized by a closed circle are concentrated primarily along the northeastern boundary of the area.

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Publication Authorized by the Director, U. S. Geological Survey

MARK A. MELTON

MULTIPLE PLEISTOCENE GLACIATION OF THE WHITE MOUNTAINS, APACHE COUNTY, ARIZONA

Abstract: Field evidence indicates two episodes of glaciation in the White Mountains, separated by a considerable time gap represented by deep weathering of the older glacial material, stream entrenchment, and outpouring of basalt and cinders over the area. The snow-level depression at the time of the younger glaciation was 2700 ± 640 feet below the estimated, modern snow level.

The White Mountains area, east and northeast of Baldy Peak, the highest point, contains evidence of multiple Pleistocene glaciation that has not been described in the literature. The extent of the glacial deposits is not yet known, and further work is under way to unravel the Pleistocene history of the area.

The highest summit of Baldy Peak (Lat. 33° 54′N., Long. 109° 34′W., elevation 11,490 feet) lies on the divide between the Little Colorado River and the Black-Salt-Gila River system (Fig. 1). East and northeast of Baldy, for 8–10 miles, is a relatively gentle, rolling, plateaulike surface at 9000 to 9300 feet elevation. This area is composed of flat-lying andesitic and basaltic lavas and bedded tuffs, whereas, near Baldy Peak, stocks of dacite porphyry support steep-sided hills. The major drainage is the West Fork of the Little Colorado River, which has eroded a canyon back into the plateau. The surfaces of the lava plateau are covered with grass; spruce and fir grow on the hill-slopes as isolated patches at lower altitudes but continuous higher in the mountains. The broad, high, plateaulike hypsographic characteristics of the White Mountains undoubtedly are effective in lowering the mean temperatures of the region, and perhaps in producing greater precipitation than would be expected from geographic position and elevation.

The evidence for an older glaciation is a deposit of till in a railroad cut approximately 1.2 miles south of Sheeps Crossing on State Highway 273. The outcrop is at an elevation of about 9300 feet, about 5 miles northeast of Baldy Peak. The exposure is 6 feet high, weathered to the base, and is capped by a well-developed, yellow-brown podzolic soil or planosol (E. A. FitzPatrick, field identification). Up the hill and back from the outcrop basalt flows and a red cinder deposit overlie the till; the contact is obscured by black soil wash.

The till contains cobbles and boulders of volcanic rocks of several types. Most notable are boulders up to 8 feet in diameter of porphyritic dacite, which were evidently brought from outside the immediate area, as the nearest known outcrop of dacite is about 3 miles away. The topography is too gentle for transport of the largest boulders either by streams or in landslides.

The situation of the till in an open meadow, several miles from the foot of the steeper part of the mountains, suggests deposition by a piedmont glacier that moved northeastward out of the higher part of the White Mountains. The degree of soil development, depth of weathering, and decay of the contained rocks date the till as pre-Wisconsin. After development of the soil on the till, the area was trenched by stream erosion and then covered by basalt flows.

The valley of the West Fork of the Little Colorado River above Sheeps Crossing was glaciated once and probably twice in Wisconsin time after dissection of the basalt flows; it is straight, lacks interlocking spurs, and has a remarkably broad floor and low gradient for its proximity to a high mountain area. The sides of the valley are altered by creep and runoff. About 2 miles above Sheeps Crossing is a deposit of boulders in the valley that appears to be a small, stream-eroded, recessional moraine. A striated cobble and many cobbles and boulders that appeared to be glacially abraded were found. Similar boulder deposits occur farther upstream.

Higher in the valley are two boggy meadows. Near the middle of the higher meadow, parallel to the length of the valley, is a string of boulders 2–4 feet in diameter which is prob-

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ably a medial moraine. Below the meadow an extensive deposit of unsorted boulders forms a terminal or large recessional moraine. The rocks are too varied to have been derived locally from the hill slopes by creep or frost action. The boulders on the surface of the moraine are not greatly weathered. Andesite and basalt boulders on the moraine show most extensive weathering—the weathered rind extends inward for about 2 inches. More silicic rocks are scarcely weathered at all. Higher in the same valley, about 3½ miles above Sheeps Crossing, a lateral moraine contains erratics up to 15 feet in diameter; these are of varying rocks types and could not have been derived from the immediate hillside. Closed depressions were seen in this lateral moraine.

In the headwaters of the valley, on the northeast side of the mountains, are two cirques at elevations of 10,600–10,800 feet. The short valleys from these two cirques join and may thus explain the origin of the medial moraine in the meadow.

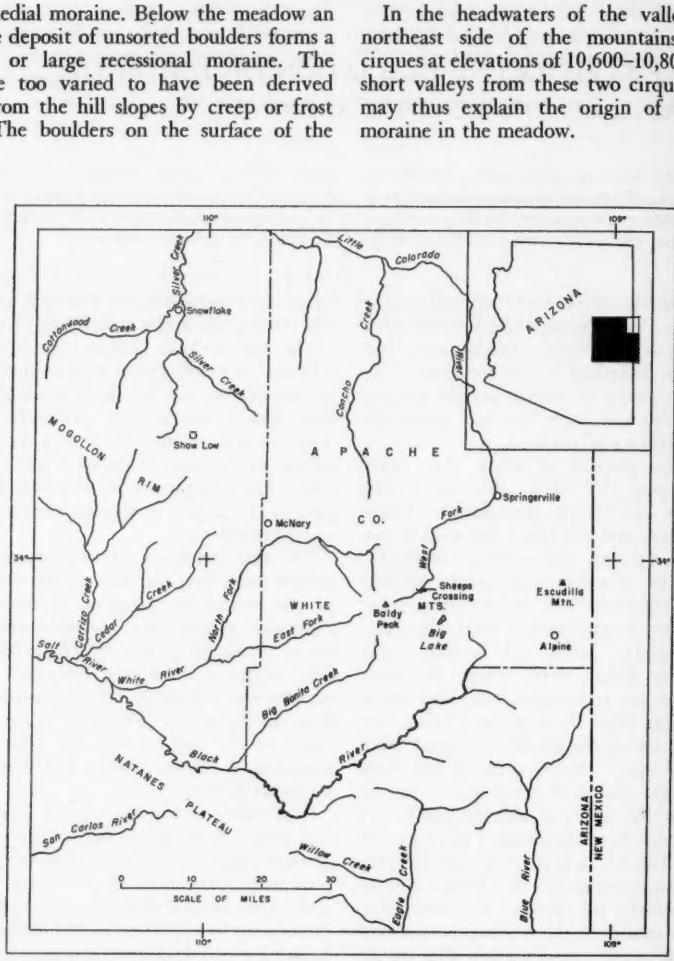

Figure 1. Drainage map and index map of White Mountains area, Arizona

The extent of the valley glacier that occupied the West Fork of the Little Colorado River is not known. The topography suggests that it must have extended at least 4 miles from the cirques. A large terminal moraine may have been deposited below Sheeps Crossing but since removed by fluvial erosion, because below Sheeps Crossing the valley becomes canyonlike and is evidently either greatly modified by or entirely of fluvial origin.

The White Mountains are not high enough

to intercept the modern orographic snow level. The elevation of the orographic snow level today is entirely hypothetical, but is approximately the elevation that would have a mean annual temperature of 25° F.; this can be estimated from climatological records. Assuming sufficiently high mountains, the 25°F isotherm would be at 13,400 feet, ±540 feet (standard error of estimate). Considering the cirque elevations, the indicated snow-level depression is therefore 2700±640 feet for the time of formation of the cirques. This compares favorably with previous estimates of Wisconsin snow-level depression for this region of about 800 m (2600 feet) (Charlesworth, v. 2, p. 653).

Acknowledgments

The author is indebted to Mr. T. L. Smiley, University of Arizona, for suggesting the possibility of finding a cirque on Baldy Peak; to Dr. Paul E. Damon, University of Arizona, for identifying volcanic rocks and making a number of helpful suggestions; to Dr. E. A. Fitz-Patrick, University of Aberdeen, Scotland, for field discussions and identifying the soils of the area; and to Dr. Paul MacClintock, Princeton University, for critically reading the manuscript. Field travel was financed by the Utilization of Arid Lands Program of the University of Arizona, supported in part by the Rockefeller Foundation.

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University of Arizona, Tucson, Ariz.
Manuscript Received by the Secretary of the Society, November 28, 1960
Contribution No. 40 of Arid Lands Program at the University of Arizona

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GLENN R. SCOTT

LATE WISCONSIN AGE OF TERRACE ALLUVIUM ALONG THE NORTH LOUP RIVER, CENTRAL NEBRASKA: A REVISION

Abstract: A radiocarbon age of 10,500 ± 250 years B.P. (Before Present) from shell material near the base of a terrace along the North Loup River, Howard County, Nebraska, dates the alluvial sequence of silts and soils as late Wisconsin. The alluvium containing the shell material previously was reported as Kansan and Yarmouth in age.

New radiocarbon dates, supplemented by stratigraphic and paleontologic evidence, indicate that an alluvial sequence along the Loup rivers in central Nebraska is much younger than we previously reported (Miller and Scott, 1955). In the earlier report we described the sequence of alluvial deposits and soils along the North Loup and Middle Loup rivers in parts of Valley and Howard counties, Nebraska. This sequence is especially well exposed in secs. 14 and 15, T. 15 N., R. 11 W., in Coopers Canyon, a ravine cut into a high terrace along the North Loup River near Elba in Howard County.

A white sand at the base of the alluvial sequence was correlated with the Grand Island of Kansan age, and an overlying alluvial silt, on which was developed a dark-gray gley soil, was correlated with the Sappa formation of late Kansan age (Miller and Scott, 1955, p. 1433). Other overlying alluvial silts and their respective soils were believed to be of Wisconsin or Recent age.

Correlation of the silt with the Sappa formation was based partly on its similarity to typical Sappa formation elsewhere, but primarily on a molluscan fauna consisting of 46 species collected from the silt and gley soil (Miller and Scott, 1955, p. 1443). This fauna includes a snail identified by Dwight W. Taylor, U. S. Geological Survey, as *Promenetus pearlettei* (Leonard), which is considered diagnostic of deposits of Kansan through Illinoian age (Taylor, 1954, p. 3). This one species thus indicated a Yarmouth and Kansan age for the gley soil and underlying deposits.

The alluvial silts and soils that constitute the overlying terrace deposits were considered to be Wisconsin and Recent in age. We assumed that the silts were deposited during glacial substages and that the soils were formed during interstadials and we tentatively correlated the lower three cycles of alluvial silts with the Iowan(?) to Tazewell(?), Cary(?), and Mankato(?) substages respectively (Miller and Scott, 1955, p. 1433, 1437). Two cycles of silt and soil overlying these were correlated with the Recent epoch.

Radiocarbon, stratigraphic, and paleontologic information that became available in 1959 suggested that these correlations were not valid. Analyses of shell material collected by J. C. Brice, U. S. Geological Survey, from beneath the gley soil gave a radiocarbon date of 10,500±250 years B.P. (W-752) (Meyer Rubin, U. S. Geological Survey, written communication, June 4, 1959; Rubin and Alexander, 1960, p. 151).

E. C. Reed and V. H. Dreeszen of the Nebraska State Geological Survey also pointed out, after a preliminary examination of well records, that the white sand we called Grand Island formation was probably an alluvial deposit of Wisconsin age. Consequently, the material we correlated with the Sappa was not the Sappa, but it and all overlying alluvial deposits were of Wisconsin age (Reed and Dreeszen, oral communication, October 1959).

After receiving this radiocarbon and stratigraphic information we requested a restudy of the fossil *Promenetus pearlettei* (Leonard). Dwight W. Taylor, U. S. Geological Survey, modified his earlier identification as follows:

"Some of the fossil *Promenetus* from zones 3 and 4 of the Elba collection from the Coopers Canyon gley soil have moderately strong riblets, rare or absent in Recent shells. None of the fossils have the very strong riblets which occur in some (but not all) of the Pliocene to middle Pleistocene samples. It would be better judgment to refer to these Elba

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shells as *Promenetus* cf. *P. exacuous*, or *P. kansasensis*, rather than identifying them precisely. These shells do not give evidence helpful for age determination."

(Written communication, October 22, 1959.)

In October 1959 we collected a sample of shell material for an additional radiocarbon age determination from the lower 4 inches of the tan silt below the gley soil and from the upper 8 inches of white sand underlying the silt. The mollusks were concentrated in a pocket that extended from the sand into the silt. The shells were cleaned in a supersonic transducer, each shell was broken, and by examination under a reading glass absolutely clean pieces were selected. The date determined for this material was 10,850 ±300 years B.P. (W-919) (Meyer Rubin, U. S. Geological Survey, written communication, December 7, 1960). As a general rule, shell material can be as much as 2000 years younger than the radiocarbon age determination (Deevey and others, 1954); consequently the shells may be as young as 8550 years.

In 1955 we correlated a soil (Miller and Scott, 1955, p. 1433) in the high terrace deposits with the Brady soil of Schultz and Stout (1948, p. 570) because of its comparable degree of development. The radiocarbon date of 10,850 years implies that all the alluvium and soils in the high terrace, including the Brady(?) soil, as well as deposits we originally thought were Kansan and Yarmouth in age, are younger than the Two Creeks interstadial. It seems likely, therefore, that the Brady soil of Schultz and Stout is younger than 10,000 years. This inference is supported by an age of 9160 ±250 years B.P. (W-234) determined for the Brady soil of Schultz and Stout at the type locality of the Bignell loess (Rubin and Suess, 1956, p. 443–444).

A further somewhat unexpected implication of the radiocarbon dates is that there were four additional soil-forming intervals in Recent time following development of the Brady(?) soil. Of these, the Ord soil (Miller and Scott, 1955, p. 1438)—the first soil above the Brady(?) soil—is most likely correlative with the altithermal soil of Antevs (1948, p. 178). The Elba, Turtle Creek, and Arcadia soils (Miller and Scott, 1955, p. 1442, 1438–1439) are late Recent in age.

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U. S. Geological Survey, Federal Center, Denver, Colo.

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S. R. HART

MINIMUM AGE OF THE MIDDLE SILURIAN IN NEW BRUNSWICK BASED ON K-Ar METHOD

Abstract: Five stratigraphically dated granitic bodies in New Brunswick have a K-Ar age between 380 and 400 m.y. Four of the granites are post-Middle Silurian and pre-Pennsylvanian in age. The fifth granite body is post-Middle Silurian and pre-Late Devonian. This indicates a minimum age of 380 m.y. for Middle Silurian time in New Brunswick.

This report contains a few K-Ar dates on samples from granitic rocks of reasonably well documented stratigraphic age from the Province of New Brunswick, Canada (Fig. 1). The study was part of a broader investigation of the geology of New Brunswick, in which the senior author collaborated with J. C. Smith, Chief Geologist, New Brunswick Department of Lands and Mines, and Dr. G. S. MacKenzie, Chairman, Department of Geology, University of New Brunswick. Analyses were made in the geochronology laboratory at Massachusetts Institute of Technology, supported by research grants to P. M. Hurley from the Division of Research, Atomic Energy Commission.

The Central Highlands of New Brunswick stretch southwest from Bathurst across the Province into Maine. This upland region is underlain by a broad belt of Paleozoic sedimentary, metamorphic, and volcanic rocks, cut by large granitic complexes of probable Devonian age. Narrow to broad aureoles of thermal metamorphism surround the granitic complexes. Unconformably overlying the rocks mentioned above is a nearly flat-lying cover of Carboniferous sedimentary rocks.

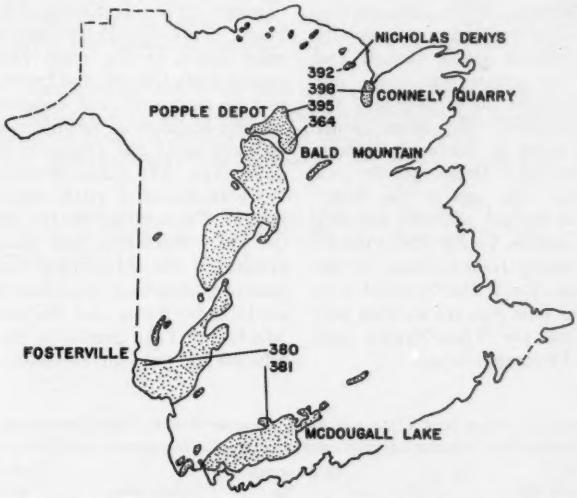

Figure 1. Distribution and age of granitic rocks in New Brunswick
Scale 1 inch equals 72 miles

Table 1 describes the samples studied. The

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TABLE 1. SAMPLE LOCATIONS AND DESCRIPTIONS

M.I.T. Sample No.	Locality	Lat. and Long.	Description
B3076	Connely Quarry, Bathurst	Lat. 47°34′N. Long. 65°40′W.	Pink biotite granite, containing phenocrysts of plagioclase
B3806*	Popple Depot, Northumberland Co.	Lat. 47°20′N. Long. 65°29′W.	Gray biotite granite
B3807	Nicholas Denys Glouchester Co.	Lat. 47°42′N. Long. 65°53′W.	Gray fine-grained biotite granite
B3833†	Bald Mountain, Northumberland Co.	Lat. 47°12′N. Long. 65°28′W.	Gray biotite granite highly altered and chloritized
B3882	McDougall Lake, Charlotte Co.	Lat. 47°27′N. Long. 66°35′W.	Pink biotite granite, containing phenocrysts of plagioclase

* Sample collected by R. Louden, University of Toronto
† Sample collected by F. D. Anderson, Geological Survey of Canada

first four samples are from the Bathurst area in northern New Brunswick. Sample B3807 is intrusive into fossiliferous Middle Silurian sedimentary rocks (Skinner, 1953). The granitic stock from which B3076 is taken is overlain by flat-lying Pennsylvanian strata (Smith and Skinner, 1958). This granite was locally unroofed and incised before the deposition of the Pennsylvanian sediments. The Bonaventure formation found north of Bathurst contains pebbles and cobbles of a similar granite in a conglomeratic zone. The age of the Bonaventure, because of the lack of fossils, has long been a subject of dispute. Young (1911) placed it, on the basis of stratigraphic evidence, in the late Devonian, and Alcock (1935) dated it as Carboniferous. The four granites are thus post Middle Silurian and pre-Pennsylvanian (and possibly pre-Late Devonian) in age.

In southwestern New Brunswick the McDougall Lake granite is intrusive into Silurian (Mascarene) sedimentary and volcanic rocks (Tupper, 1959; MacKenzie, 1951; Alcock and Perry, 1945). The Perry formation is found 3 miles south of the main McDougall Lake granite body (Alcock and Perry, 1945). It consists of sandstones and conglomerates, which contain boulders of fossiliferous Silurian sedimentary rocks and granite of the McDougall Lake type. The Late Devonian age of the Perry formation is established from its plant remains (Alcock and Perry, 1945). In Maine, the Perry formation rests unconformably on granite of the McDougall Lake type, and granite boulders are abundant in its conglomeratic beds (Bastin and Williams, 1914). The McDougall Lake granite is thus post-Middle Silurian and pre-Late Devonian in age.

TABLE 2. ANALYTICAL DATA FOR K-AR AGES OF BIOTITE, NEW BRUNSWICK

M.I.T. Sample No.	% K	Air Ar correction %	Ar_{40}^*/K_{40}	K-Ar age (m.y.)
B3076	7.00, 7.05 Av. = 7.02	Unsatisfactory	0.0260	398±12†
B3806	7.03, 7.05 Av. = 7.04	9.1	0.0258	395±15
B3807	4.84, 4.87 Av. = 4.86	0.5	0.0255	392±12
B3833	4.85, 4.81, 4.54 Av. = 4.73	8.6	0.0234	364±15
B3882	5.00, 5.02 Av. = 5.01	2.6	0.0247	381±11

* Radiogenic argon-40
† 398 m.y. is a maximum value only

Argon analyses were made by isotope dilution, using enriched Ar^{38} as a spike and Ar^{36} as a measure of air-argon contamination. Reproducibility of an argon analysis is normally better than 3 per cent in this age range. Potassium analyses were made using a Perkin-Elmer Model 146 flame-photometer with a lithium internal standard. Each potassium measurement is the average of at least two flame-photometer determinations, and the reproducibility of a single measurement is approximately 3 per cent. Decay constants used are $\lambda_e = 0.585 \times 10^{-10}$ yr.$^{-1}$; $\lambda = 5.30 \times 10^{-10}$ yr.$^{-1}$; and $K^{40} = 1.22 \times 10^{-4}$ g/g of K. The analytical results are shown in Table 2.

A date of 380–390 million years can thus be assigned to geologic time somewhere between post-Middle Silurian and pre-Late Devonian in southern New Brunswick, and post-Middle Silurian and pre-Pennsylvanian in northern New Brunswick. In all probability these granites are Early Devonian in age. This indicates a minimum age for Middle Silurian time (380 m.y.) and forces an appreciable extension of Holmes' B time scale (1947).

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M.I.T. Age Studies No. 19

Proceedings of the Society

Rocky Mountain Association of Geologists Field Conference

A field conference to be held on September 7 and 8 will study the lower and middle Paleozoic section in the Salida, Monarch, Ouray, Silverton, and Durango areas. A technical session will be held in Salida on the night of September 6. Additional information may be obtained from Denzil W. Bergman, Conference Chairman, 1600 Ogden Street, Denver 5, Colorado

Proceedings of the 14th Annual Meeting of the Rocky Mountain Section

University of Wyoming, Laramie, Wyoming, May 11–13, 1961

John de la Montagne, *Secretary*

The fourteenth annual meeting of the Rocky Mountain Section of The Geological Society of America was held in Laramie, Wyoming, May 11–13. Technical sessions were held the first two days, but inclement weather forced the cancellation of the field trips on May 13.

There were 151 official registrants, 7 registrants by compliment, and 52 student and other nonpaying registrants.

ANNUAL LUNCHEON AND BUSINESS MEETING

The fourteenth annual business meeting of the Rocky Mountain Section of The Geological Society of America was held in the private dining room of the Knight Hall Cafeteria on the campus of the University of Wyoming, May 12. The meeting was called to order at 12:45 p.m. by Chairman William R. Keefer. Eighty-seven Members and Fellows, and two guests (AGI Visiting International Scientists) were present.

The minutes of the Rapid City meeting were read and approved as read. The financial report showed a balance of $1026.58 of which $927.79 represented savings before interest at the Midland Federal Savings and Loan Association, Denver, and $98.79 remained in the First National Bank of Bozeman, Montana, as an operating fund checking account. The approval of the Financial Report was moved and seconded by Allen Agnew and Warren Thompson. An auditing committee consisting of Keith Rigby and David Clark reported that the Interim Financial Report as of May 6, 1961, had been inspected and approved.

The Chairman read the report of the Ballot committee showing that 174 valid votes were cast for the incoming officers who were then introduced, as follows:

Dr. J. K. Rigby, Brigham Young University, Provo, Utah, *Chairman*
Dr. David Clark, Brigham Young University, Provo, Utah, *Vice-Chairman*

In accordance with the resolution adopted at the Missoula meeting, 1959, the present Secretary will continue to serve in that capacity.

There was no old business.

Chairman Keefer commented favorably upon the newly instigated division of labor for the officers in which the Chairman and Vice-Chairman are most logically heads of the local committee and the Secretary remains permanently in charge of official correspondence, manuscript review, and finances but is not necessarily in residence at the meeting sites.

The Chairman announced the invitation from Dr. Stuart Northrop on behalf of the University of New Mexico, Albuquerque, for the section to meet on that campus during the Spring, 1963. A motion to accept this invitation was proposed and seconded by Warren Thompson and M. Van Couvering.

The Secretary stated that, in practice, the Section was not complying with existing Bylaws in providing for a three-man Nominating Committee for officers. Accordingly, an amendment to the Bylaws was moved by G. D. Robinson, and seconded by Frank Swenson. The amendment reads as follows: Section 1. *Method of electing officers.* The incoming officers of the Section each year shall appoint a nominating committee of three, at least one of whom shall be a past officer of the Section. (The remainder of the paragraph is unchanged). The amendment was passed unanimously.

The Chairman introduced the following officers of The Geological Society of America:

Dr. Thomas Nolan, *President*
Dr. M. K. Hubbert, *Vice-President*
Dr. Frederick Betz, Jr., *Secretary*
Dr. E. B. Eckel, *Councilor*

Doctor Nolan commented that this particular sectional meeting resembled the early national meetings in spirit and numbers present, and outlined problems of the Society concerning the pending move to new quarters, the cost of financial management, and the reorganization of publication policy.

Doctor Hubbert described the "Fellowship problem" and brought the Section up to date on current developments regarding it.

Doctor Betz elaborated on other aspects of Fellowship candidacy and described the problems of growth of the national meeting. He stated that sectional meetings will continue to rise in stature and suggested that the sections begin to think in terms of systematically sponsoring national meetings on a rotational schedule.

Doctor Eckel congratulated the Section on its growth and strength and complimented the local committee for its work in arranging for the 1961 meeting at Laramie.

The meeting was adjourned at 1:30.

ANNUAL DINNER

A total of 151 members, wives, and guests attended the Annual Dinner held in the dining room of Knight Hall Cafeteria. Drs. Jean Goguel, Ingenieur General des Mines, Paris, France, and Hisashi Kuno, Professor of Petrology, Geological Institute, University of Tokyo, Japan, addressed the assembly on the subjects of alpine tectonics and pumice flow, respectively.

FIELD TRIPS

A Spring blizzard caused the cancellation of the planned field trip to places of interest in the Laramie Basin area for May 13.

SCIENTIFIC SESSIONS

The following sessions were held on May 11 and 12:

General Session

Cochairmen: Allen F. Agnew and Horace D. Thomas

1. John Chronic* and C. S. Ferris, Jr.: *Lower Paleozoic outlier in southeastern Wyoming* 9:15
2. J. E. Case* and H. R. Joesting: *Northeast-trending Precambrian structures in the central Colorado Plateau* 9:35
3. R. P. Sheldon, E. K. Maughan*, and E. R. Cressman: *Sedimentation in Wyoming and adjacent areas during Leonard (Permian) time* 10:00
4. R. S. Houston* and M. E. McCallum: *Mullen Creek–Nash Fork shear zone, Medicine Bow Mountains, southeastern Wyoming* 10:20
5. Wm. Lee Stokes: *Geologic map of Utah* 10:40
6. J. L. Weitz: *State line fault and other geofallacies* 10:55
7. E. L. Tullis: *Basic Science curriculum in geology* 11:15

*Indicates speaker

Stratigraphy and Pleistocene Geology

Cochairmen: J. Keith Rigby and Robert J. Weimer

1. P. E. Soister: *Stratigraphy of Wind River formation (lower Eocene), Gas Hills uranium district, Wind River Basin, Wyoming* 1:30
2. R. H. De Voto: *Tertiary stratigraphy of South Park, Colorado* 1:50
3. Yazdan Mollazal: *Ely limestone in parts of Utah and Nevada* 2:15
4. T. C. Hiestand: *Reservoir rock bodies and enclosing rocks mapped in Cretaceous formations, Denver and Powder River basins* 2:35
5. J. Stewart Williams: *Lake Bonneville group in Cache Valley, Utah* 3:00
6. B. M. Irvine: *Geologic investigation of landslides in the Ralston Creek area, Jefferson County, Colorado* 3:20
7. M. E. McCallum: *Glaciation of Libby Creek Canyon, east flank of Medicine Bow Mountains, southeastern Wyoming* 3:45

Areal Geology

Cochairmen: William H. Wilson and William J. McMannis

1. H. J. Winkler: *Geology of the Scipio Pass area, Juab and Millard counties, Utah* 1:30
2. J. W. Guyton: *Late Cretaceous-early Tertiary history of the Lost Soldier area, Wyoming* 1:50
3. C. N. Murdock: *Geology of the Weston Canyon area, Bannock Range, Idaho* 2:05
4. C. R. Dunrud: *Volcanic and sedimentary rocks in the Jack Creek area, Park County, Wyoming* 2:25
5. P. B. Schmidt: *Geology of the State Bridge area, Colorado* 2:45
6. D. B. Andretta: *Geology of the Moose Creek stock, Highland Mountains, Montana* 3:10
7. R. W. Scott: *Geology of the Rabbit Ears Pass area, Jackson and Grand counties, Colorado* 3:25

Stratigraphy and Engineering Geology

Cochairmen: William H. Curry and Bruce F. Curtis

1. D. W. Bolyard* and John Chronic: *Pennsylvanian stratigraphy in Sangre de Cristo Mountains, Colorado* 9:00
2. Bronson Stringham: *Precambrian stratigraphy of the Grouse Creek Mountains, northwestern Utah* 9:25
3. G. B. Robinson, Jr.: *Leonardian Pequop formation (Permian) in central Pequop Mountains, Nevada* 9:50
4. K. A. Hodgkinson: *Permian stratigraphy of part of northwestern Utah* 10:10
5. J. V. A. Sharp: *Stratigraphy of the Dakota, Mowry, and Frontier formations of the Piceance basin, Colorado* 10:30
6. D. L. Warner: *Stratigraphy of the lower portion of the Mesaverde formation, southeastern Piceance basin, Colorado* 10:55
7. W. F. Sherman: *Engineering geology of the Cody Tunnel project, Wyoming* 11:15

*Indicates speaker

Mineralogy and Petrology

Cochairmen: Paul K. Sims and John P. Wehrenberg

1. Ming-Shan Sun: *Oriented growths of cryptomelane in sylvite, Carlsbad, New Mexico* 9:00
2. B. J. Anderson* and J. A. Whelan: *Nickel-bearing minerals in the southern Oquirrh Mountains, Utah* 9:15
3. D. L. Hoover: *Geology of the Lark mine and vicinity, San Juan County, Colorado* 9:30
4. G. B. De Koster: *Petrology of carbonate rocks in the Goose Egg formation, eastern Big Horn Basin, Wyoming* 9:55
5. J. T. O'Connor: *Metamorphic rocks of the staurolite-almandine subfacies from southern Buckhorn Mountain, Larimer County, Colorado* 10:20
6. Howard Stensrud: *Lake Owens mafic complex, Medicine Bow Mountains, Wyoming* 10:45
7. R. E. Boyer* and H. D. King: *Petrogenetic significance of heavy minerals in the San Isabel batholith, Wet Mountains, Colorado* 10:55
8. W. H. Taubeneck: *Zircons in granitic rocks near Cornucopia, Wallowa Mountains, northeastern Oregon* 11:15

Geochemistry and Precambrian Geology

Cochairmen: A. L. Slaughter and R. M. Hutchinson

1. R. T. Chew III and H. T. Mitten*: *Geochemical prospecting in northwest Montana* 1:30
2. R. E. Riecker: *Geologic interpretations of hydrocarbon fluorescence under ultraviolet light* 1:55
3. M. Koksoy* and L. W. LeRoy: *Trace elements in lichens—a possible geochemical exploration tool* 2:15
4. J. P. Wehrenberg* and Arnold Silverman: *Diffusion of base metals at Gilman, Colorado* 2:35
5. D. N. Stevens* and M. A. Klugman: *Trace-element dispersion in quartzite wall rock at the Phillips mine, Park County, Colorado* 2:55
6. S. Ramarathnam: *Geology and petrology of the southern part of the Laramie anorthosite, Albany County, Wyoming* 3:20
7. T. H. Murray, Jr.: *Anorthosite and the ilmenite-magnetite deposits of a portion of the Laramie anorthosite, Albany County, Wyoming* 3:40

Structure and Paleontology

Cochairmen: J. Stewart Williams and Reuben J. Ross, Jr.

1. M. H. Manghnani* and John Hower: *Structural significance of a gravity profile in the Bitterroot Valley, Ravalli County, Montana* 1:30
2. J. K. Rigby* and E. L. Fitzgerald: *Medial and late Paleozoic structural history of western Alberta and eastern British Columbia* 1:45
3. L. W. McGrew: *Structure of Cenozoic rocks along the northwestern margin of the Julesburg Basin, southeastern Wyoming* 2:05

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*Indicates speaker

4. H. E. Holt: *Structural features of the Blue River valley, Colorado* 2:25
5. F. A. Donath and R. B. Parker*: *Folds and folding* 2:45
6. R. B. Hoy*, R. M. Foose, and B. J. O'Neill, Jr.: *Structure of the Winnfield Salt Dome, Winn Parish, Louisiana* 3:10
7. D. W. Trexler: *Coccolithophorid assemblages from the Benton and Niobrara groups, Canon City area, Colorado* 3:30
8. Heinrich Toots: *Burrowing organisms and their effect on sediments* 3:50
9. K. R. Newman: *Microfossil correlations of Upper Cretaceous and Paleocene formations, Sand Wash and Piceance basins, northwestern Colorado* 4:10

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By Title

E. M. Schell: *Geology of the Marietta mine, Broadwater County, Montana*
G. E. Sorensen, Jr.: *Geology of the Thousand Springs Valley area, Madison and Teton counties, Idaho*
H. T. Stearns: *Unusual glassy basaltic dikes at Copperfield, Oregon*
H. T. Stearns: *Sill at the base of the Columbia River basalts near Robinette, Oregon*
W. H. Taubeneck: *Zeolites in granitic rocks from northeastern Oregon*

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